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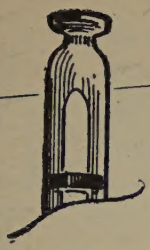

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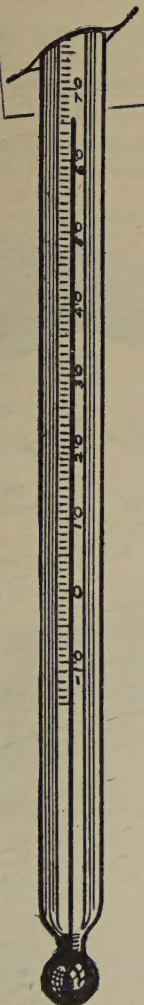

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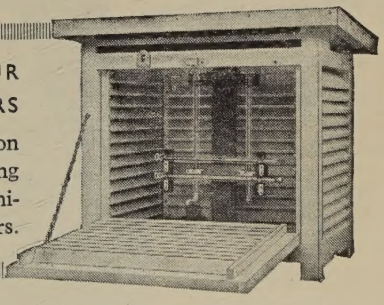

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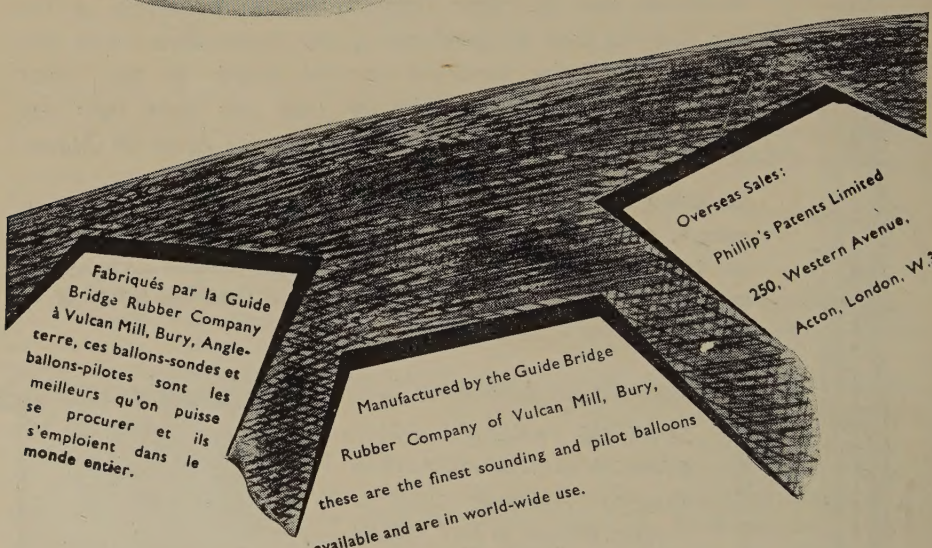

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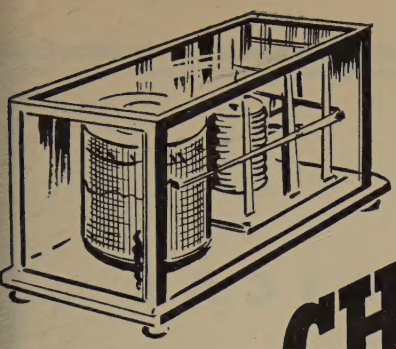

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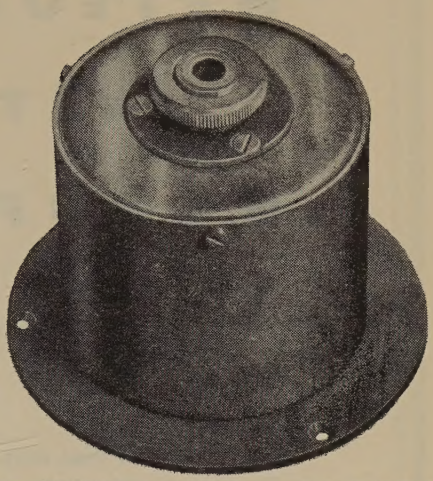

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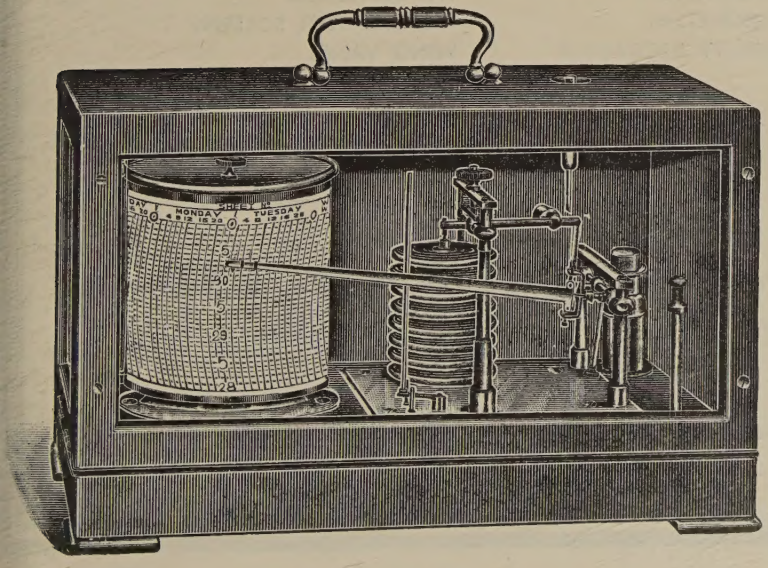

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Some recent measurements of humidity from aircraft up to heights of about 50,000 ft over southern England

By R. J. MURGATROYD, P. GOLDSMITH and W. E. H. HOLLINGS
Meteorological Research Flight, Farnborough, Hants.

(Manuscript received 25 May 1955)

SUMMARY

The humidity of the air was measured with a pressurized Dobson-Brewer hygrometer in a Canberra aircraft up to heights of about 50,000 ft over southern England on 35 occasions during 1954. The frost point decreased with increasing height in the stratosphere and usually approached a constant value of − 115°F to − 120°F a few thousand feet above the tropopause.

1. INTRODUCTION

The humidity measurements in the upper troposphere and lower stratosphere made in aircraft of the Meteorological Research Flight at Farnborough and the High Altitude Flight at Boscombe Down from 1943 to 1950, which have been presented by Bannon, Frith and Shellard (1952), extended up to a height of about 40,000 ft, the ceiling of the piston-engined aircraft then available. Late in 1953 a Canberra jet aircraft was obtained and it then became possible to extend the ceiling of these measurements to about 50,000 ft. 35 ascents were made for this purpose in 1954 and are described below.

2. INSTRUMENTATION

The pressurized form of the aircraft frost-point hygrometer described by Brewer, Cwilong and Dobson (1948) was used for these measurements. This instrument requires experienced operators to obtain measurements of frost points below about − 100°F, and when they are below − 120°F the frost deposit becomes difficult to recognize. In order to extend its range and, incidentally, guard against possible small leaks of cabin air into the instrument, measurements were also made on some occasions using air which had passed through the jet engine compressor, a method due to Goldsmith (1955). Good agreement was obtained between the frost points measured in this way and those obtained using outside air directly. Temperatures were measured using Meteorological Office flat-plate resistance thermometers with balanced-bridge indicators, the readings being corrected for airspeed.

3. THE OBSERVATIONS

All the readings of temperature and frost points obtained have been listed in Table 1. The mean values at the heights given have been plotted to obtain the mean curves shown in Fig. 1. In addition mean temperature and frost-point values have been plotted at heights above and below the tropopause in the two groups shown in Fig. 2. The group listed as 'high tropopause' cases refers to ascents when the tropopause was above 35,000 ft and that of the 'low tropopause' cases to occasions when it was below that level.

4. DISCUSSION

The preliminary results from this set of measurements are :

(1) The frost point over southern England decreases with height above the tropopause and tends to approach a constant value, usually between − 115°F and − 120°F, a few thousand feet above this level.

TABLE 1. TEMPERATURE AND FROST POINT (IN BRACKETS) °F, INDIVIDUAL ASCENTS (1954)

I.C.A.N. Height (ft)	Pressure (mb)	Dates: 19 Jan.	22 Feb.	24 Feb.	25 Feb.	5 Mar.	10 Mar.	11 Mar.	22 Mar.	23
		Times (GMT): 1335	1555	1110	1115	1505	1220	1515	1505	10
5,000	843	+49 (–)	+42 (+27)	+34 (+21)	+42 (+17)	+28 (+15)	+41(+21)	–	+42 (+25)	+35 5
10,000	697	+34 (+10)	+33 (+12)	+17 (+04)	+12 (−04)	–	+22 (−25)	–	+33 (−02)	+27 7
15,000	572	+18 (−21)	– (−14)	−03 (−19)	+01 (−52)	−09 (−67)	–	+07 (−28)	+14 (−40)	+06 6
20,000	465	−03 (−46)	−08 (−29)	−25 (−42)	−21 (−64)	–	−14 (−41)	–	−07 (−19)	−18 8
25,000	376	−24 (−62)	−29 (−63)	−46 (−66)	−31 (−71)	−48 (−68)	−35 (−66)	–	−26 (−28)	−37 7
30,000	301	−50 (−63)	−50 (−57)	−54 (−88)	−53 (−81)	−58 (−77)	−52 (−74)	−50 (−81)	−52 (−53)	−57 7
32,000	274	–	−60 (−66)	–	−55 (−87)	–	−59 (−79)	−56 (−85)	−59 (−67)	
33,000	262	−65 (−75)	–	−62 (−90)	–	–	–	–	–	
34,000	250	–	−69 (−74)	–	−60 (−92)	–	−65 (−83)	−60 (−91)	−66 (−73)	
35,000	238	–	–	–	–	−71 (−97)	–	–	–	−55 5
36,000	227	−78 (−85)	−78 (−77)	−61 (−98)	−54 (−102)	–	−69 (−89)	−61 (−98)	−69 (−79)	
37,000	216	–	–	–	–	–	–	–	–	
38,000	206	−82 (−90)	−84 (−87)	−62 (−102)	−51 (−108)	−64 (−109)	−70 (−97)	−59 (−109)	−71 (−86)	
39,000	197	–	–	–	–	–	–	–	–	
40,000	187	−81 (−99)	−85 (−92)	−56 (−108)	−50 (−112)	–	−65 (−105)	−57 (−115)	−70 (−93)	−52 2
41,000	178	–	–	–	–	−61 (−112)	–	–	–	
42,000	170	−71 (−113)	−78 (−102)	−55 (−115)	−54 (−115)	–	−63 (−113)	−57 (−119)	−66 (−101)	−52 2
43,000	162	–	–	–	–	–	–	–	–	
44,000	155	−72 (−124)	−67 (−105)	−59 (−115)	−55 (−117)	−59 (−115)	−63 (−115)	−58 (−122)	−59 (−108)	−51 1
45,000	147	–	–	–	–	–	–	–	–	
46,000	140	−72 (−121)	−63 (−107)	−56 (−113)	−55 (−116)	−61 (−116)	−63 (−119)	−64 (−123)	−57 (−115)	−52 2
48,000	127	−71 (−120)	−64 (−108)	−57 (−113)	−50 (−118)	−61 (−117)	−60 (−120)	−63 (−122)	−54 (−117)	−52 2
49,000	121	–	–	–	–	−63 (−117)	–	–	–	
50,000	116	−71 (−121)	−67 (−107)	−61 (−113)	−54 (−118)	–	−62 (−120)	−68 (−122)	−59 (−120)	−55 5
51,000	111	–	–	–	–	–	–	–	–	

I.C.A.N. Height (ft)	Pressure (mb)	Dates: 25 May	28 May	10 June	11 June	16 June	24 June	28 June	29 June	1
		Times (GMT): 1345	0935	1100	0930	0940	0835	1015	0915	11
5,000	843	+45 (+38)	+55 (+35)	–	–	+49 (−01)	–	–	+40 (+33)	
10,000	697	+25 (+12)	+34 (+25)	–	–	+38 (+25)	–	–	+33 (+12)	
15,000	572	+11 (+13)	+17 (+15)	–	–	+22 (+14)	–	–	+15 (00)	
20,000	465	−04 (−09)	−02 (−03)	–	–	+04 (−11)	–	–	−05 (−34)	
25,000	376	−27 (−26)	−22 (−38)	–	–	−20 (−36)	–	–	−25 (−48)	
30,000	301	−48 (−49)	−41 (−46)	−51 (−74)	−54 (−74)	−42 (−57)	−39 (−48)	−52 (−65)	−47 (−62)	−45 5
32,000	274	−58 (−62)	−49 (−56)	–	–	–	–	–	−55 (−68)	
33,000	262	–	–	–	–	–	–	–	–	
34,000	250	−64 (−76)	−57 (−62)	–	–	–	–	–	−57 (−84)	
35,000	238	–	–	−48 (−92)	−51 (−103)	−67 (−78)	−60 (−67)	−53 (−95)	–	−60 0
36,000	227	−67 (−86)	−61 (−81)	–	–	–	–	–	−50 (−99)	
37,000	216	–	–	–	–	–	–	–	–	
38,000	206	−61 (−90)	−64 (−85)	–	–	–	–	–	−47 (−107)	
39,000	197	–	–	–	–	–	–	–	–	
40,000	187	−62 (−101)	−71 (−91)	−46 (−113)	−44 (−115)	−56 (−104)	−76 (−90)	−42 (−108)	−44 (−112)	−54 4
41,000	178	–	–	–	–	–	–	–	–	
42,000	170	−58 (−105)	−68 (−106)	–	–	–	–	–	−44 (−117)	
43,000	162	–	–	−46 (−114)	−46 (−118)	−59 (−108)	−63 (−107)	−44 (−110)	–	−52 2
44,000	155	−55 (−114)	−58 (−112)	–	–	–	–	–	−45 (−120)	
45,000	147	−54 (−120)	–	–	–	–	–	–	–	
46,000	140	–	−54 (−113)	−45 (−118)	−48 (−120)	−60 (−111)	−60 (−117)	−47 (−111)	−45 (−120)	−52 2
48,000	127	–	−57 (−117)	−45 (−118)	−47 (−121)	−58 (−110)	−60 (−118)	−47 (−)	−47 (−120)	−53 3
49,000	121	–	–	–	–	–	–	–	–	−50 0
50,000	116	–	−59 (−119)	–	−49 (–)	–	−61 (−119)	–	−46 (−120)	
51,000	111	–	–	–	–	–	–	–	–	

TABLE 1 (*continued*)

..N. ht Pressure (mb)	5 July 1510	6 July 0950	14 July 1440	4 Aug. 1025	Dates 5 Aug. Times (GMT) 1010	10 Aug. 1100	11 Aug. 1000	17 Aug. 1100	23 Aug. 1000
00 843	–	–	+49 (+26)	+58 (+24)	–	–	+38 (+34)	–	–
00 697	–	+18 (−04)	+33 (−11)	–	–	+25 (+17)	–	–	–
00 572	–	−01 (−33)	+17 (−44)	–	–	–	–	–	–
00 465	–	−22 (−35)	−02 (−55)	+03 (−14)	–	–	–	00 (−12)	−11 (−38)
00 376	–	−40 (−59)	−18 (−36)	–	–	–	–	–	–
00 301	−45 (−85)	−49 (−83)	−38 (−56)	−44 (−57)	−35 (−52)	−51 (−59)	−47 (−53)	−42 (−67)	−50 (−77)
00 274	–	–	–	–	–	–	–	–	–
00 262	–	–	−53 (−58)	−55 (−70)	−52 (−65)	−58 (−75)	−53 (−80)	−56 (−62)	−57 (−86)
00 250	–	–	–	–	–	–	–	–	–
00 238	−41 (−104)	−42 (−102)	–	–	–	–	–	–	–
00 227	–	–	−64 (−69)	−61 (−83)	−65 (−75)	−51 (−87)	−49 (−92)	−61 (−84)	−54 (−99)
00 216	–	–	–	–	–	–	–	–	–
00 206	–	–	–	−53 (−93)	−66 (−92)	−46 (−92)	−51 (−96)	−60 (−95)	−49 (−99)
00 197	–	–	−66 (−96)	–	–	–	–	–	–
00 187	−43 (−111)	−42 (−112)	–	−49 (−98)	−69 (−73)	−45 (−99)	−51 (−103)	−52 (−103)	−48 (−105)
00 178	–	–	–	–	–	–	–	–	–
00 170	–	–	−58 (−112)	−54 (−105)	−65 (−86)	−47 (−100)	−54 (−106)	−56 (−107)	−49 (−106)
00 162	−41 (−115)	−42 (−112)	–	–	–	–	–	–	–
00 155	–	–	−55 (−116)	−56 (−110)	−67 (−90)	−48 (−108)	−54 (−111)	−59 (−113)	−46 (−110)
00 147	–	–	–	–	–	–	–	–	–
00 140	−46 (−117)	−45 (−113)	−53 (−119)	−60 (−114)	−68 (−90)	−50 (−111)	−57 (−115)	−59 (−116)	−49 (−112)
00 127	−49 (−118)	−48 (−116)	−56 (−119)	−59 (−118)	−67 (−89)	−49 (−116)	−56 (−117)	−60 (−118)	−51 (−113)
00 121	–	–	–	–	–	–	–	–	–
00 116	−49 (−)	−45 (−117)	−57 (−121)	−60 (−120)	−61 (−99)	–	−56 (−121)	−62 (−121)	−56 (−118)
00 111	–	–	–	–	–	–	–	–	–

..N. ght Pressure) (mb)	26 Aug. 1005	9 Dec. 1210	10 Dec. 1045	13 Dec. 1100	Dates 14 Dec. Times (GMT) 1100	20 Dec. 1420	21 Dec. 1420	22 Dec. 1355
00 843	–	–	–	–	+42 (−)	+40 (−03)	+43 (+31)	–
00 697	–	–	+08 (−26)	–	+31 (−37)	+27 (00)	+24 (+18)	+27 (+15)
00 572	–	–	−13 (−43)	+08 (−10)	+15 (+09)	+16 (−38)	–	+09 (00)
00 465	+05 (−19)	–	–	–	−07 (−12)	−10 (−54)	−04 (−55)	−11 (−19)
00 376	–	–	−60 (−63)	−31 (−40)	−30 (−40)	−33 (−56)	−26 (−39)	−25 (−41)
00 301	−39 (−61)	−67 (−81)	−67 (−93)	−51 (−59)	−48 (−64)	−57 (−77)	−45 (−68)	−49 (−63)
00 274	–	−65 (−93)	−66 (−104)	−60 (−74)	−55 (−69)	−65 (−80)	–	−57 (−76)
00 262	−54 (−65)	−66 (−98)	–	–	–	–	–	–
00 250	–	–	−68 (−110)	−67 (−82)	−66 (−71)	−73 (−84)	–	−66 (−75)
00 238	–	−65 (−104)	–	–	–	–	−70 (−71)	–
00 227	−66 (−73)	–	−64 (−113)	−72 (−86)	−76 (−81)	−77 (−89)	–	−76 (−91)
00 216	–	−69 (−111)	–	–	–	–	–	–
00 206	−67 (−81)	–	−69 (−116)	−69 (−97)	−82 (−91)	−81 (−98)	–	−83 (−90)
00 197	–	−68 (−112)	–	–	–	–	–	–
00 187	−66 (−85)	–	−68 (−120)	−70 (−107)	−75 (−105)	−77 (−104)	−95 (−98)	−92 (−100)
00 178	–	−69 (−112)	–	–	–	–	–	–
00 170	−64 (−93)	–	−68 (−121)	−72 (−108)	−75 (−116)	−81 (−111)	–	−97 (−109)
00 162	–	–	–	–	–	–	–	–
00 155	−59 (−105)	–	−72 (−122)	−68 (−112)	−74 (−119)	−75 (−113)	–	−84 (−114)
00 147	–	–	–	–	–	–	−82 (−111)	–
00 140	−62 (−110)	–	−71 (−122)	−70 (−116)	−75 (−123)	−80 (−115)	–	−84 (−117)
00 127	−66 (−111)	–	−75 (−122)	−75 (−115)	−76 (−123)	−73 (−117)	–	−77 (−117)
00 121	–	–	–	–	–	–	–	–
00 116	−65 (−116)	–	–	−72 (−117)	−76 (−125)	−73 (−117)	–	−73 (−119)
00 111	−66 (−118)	–	–	–	–	–	–	–

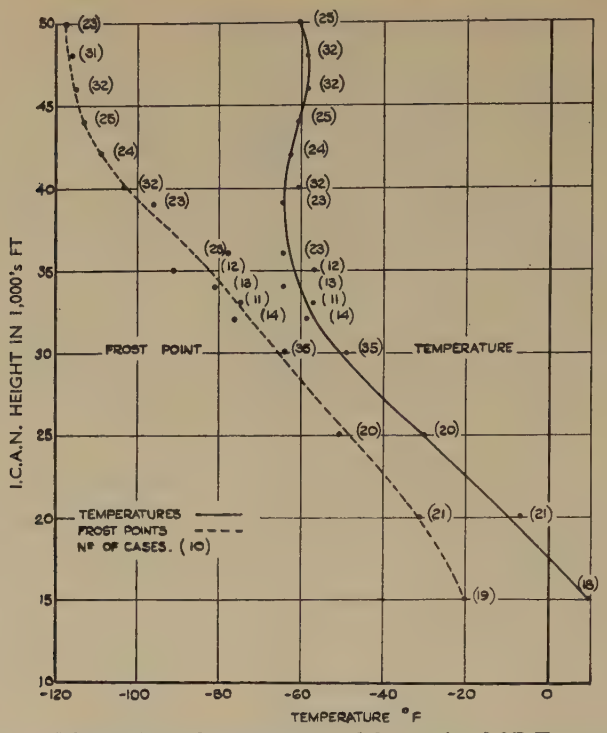

Figure 1. Mean values of temperature and frost point, M.R.F. ascents 1954.

(2) The lapse rate of frost point also decreases with height, being on the average about 4°F in the 1,000 ft above the tropopause but less than 1°F per 1,000 ft at heights of about 15,000 ft above the tropopause. Usually no discontinuity in the lapse rate of frost point was observed at the tropopause. In some cases however the change from a large lapse rate of frost point to a lower lapse rate higher up appeared to be discontinuous, suggesting a ' humidity tropopause ' a few thousand feet above the temperature tropopause. The distance between these two levels varied between 2,000 and 10,000 ft and no relationship between it and other meteorological variables has yet been noticed.

(3) Owing to aircraft unserviceability the ascents could not be evenly spaced throughout the year but readings were obtained in all four seasons. No significant difference of frost point at 50,000 ft between winter and summer has been found.

(4) When the heights are referred to the tropopause level the ' high tropopause ' cases show a lower temperature and also a lower frost point, level for level, compared with the ' low tropopause ' cases. In general, high and low values of frost points in the lowest 10,000 ft or so of the stratosphere correspond well with high and low frost-point values at the tropopause, but this correspondence disappears at higher altitudes as the frost point approaches constancy.

The accuracy of the temperatures quoted is thought to be ± 2°F, a value deduced from the variability of the airspeed correction of the flat-plate thermometer in trials made for this purpose at high speeds with the Canberra aircraft. The heights quoted may be up to 500 ft in error, a figure obtained from comparative trials between the pilot's altimeter and a radar altimeter. The frost points have a random error of ± 3°F due to uncertainties in reading the hygrometer under these conditions. An analysis of humidities measured with this instrument in the neighbourhood of high cloud by Murgatroyd and Goldsmith (1953) gave the result that the air there had a frost point on the average 3°F lower than would have been expected if the air were saturated with respect to ice. This

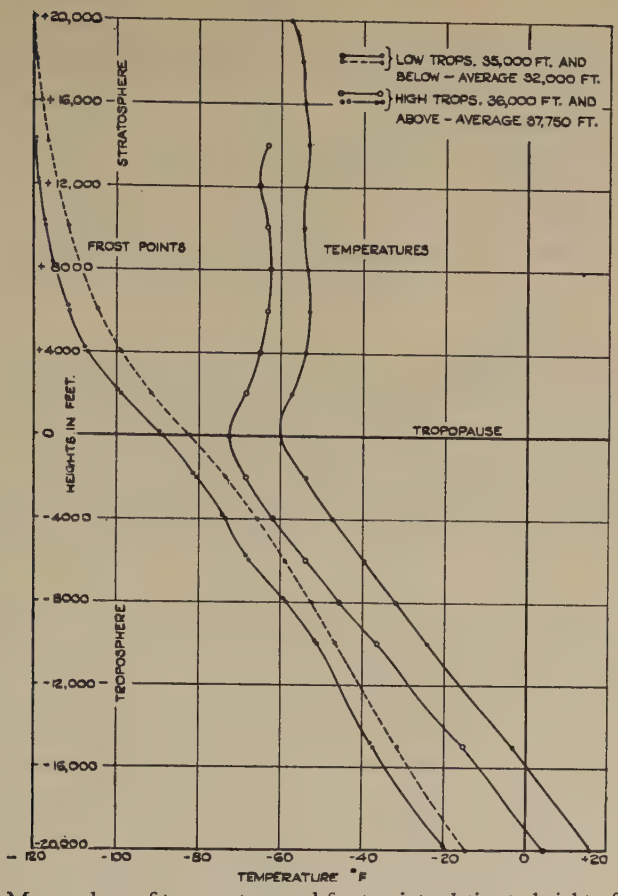

Figure 2. Mean values of temperature and frost point relative to height of tropopause.

suggests that the instrument may also have a constant error, reading somewhat low at these altitudes but it is unlikely that the error will be more than a few degrees Fahrenheit.

The general picture obtained from these measurements is therefore one of variable frost points at the tropopause tending to constancy of frost point at — 115°F or so at levels of 10,000 to 15,000 ft above it. The air at these altitudes is therefore very dry with a relative humidity with respect to ice of less than 1 per cent. The range of variation of frost-point values there is usually small although considerably higher frost-point values were observed on one occasion (5 August 1954). These values of frost point are in general agreement with those expected from the tendency observed at the greatest heights by Bannon, Frith and Shellard (1952), although they are a few degrees lower.

It is hoped to continue measurements of this type throughout 1955.

Acknowledgment

We are indebted to the Director of the Meteorological Office for permission to publish this paper.

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Brewer, A. W., Cwilong, B. M. and Dobson, G. M. B.	1948	*Proc. Phys. Soc.*, **60**, p. 52.
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551.558.1 : 551.555.3

The effects of a large island upon the trade-wind air stream

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SUMMARY

A series of four helical aircraft soundings across the island of Puerto Rico on a sea-breeze day are analyzed and interpreted. The existence of a fairly symmetrical orographic-convection cell with ascent over the island and compensating subsidence in a surrounding ring is established. The field of motion in the convection cell is deduced from the observations together with the theoretical work on flow over heated islands by Malkus and Stern (1953). The plausibility of the motion field is tested by construction of a moisture budget for part of the cross-section delineated by the soundings, and the effects of the subsidence, lifting, heating and other processes imposed upon the air stream by its passage over the island are discussed, with particular emphasis upon alterations in the upward transport of water vapour and cumulus cloud structure.

1. INTRODUCTION

Although diurnal variations in a trade-wind air mass over the ocean are extremely slight, the presence of an island or land-body introduces marked changes from night to day. The alteration in cloud patterns is particularly striking. Over Puerto Rico and even the smaller Virgin Islands, the build-up of land and orographic clouds during the morning is commonly accompanied by clearing at the coastlines and the formation of a clear ring surrounding the island. Except in the localities where the land-formed clouds may extend seaward in a ' street,' the cloud-free ring extends off-shore with a radius comparable to the island width. In the late afternoon as diurnal convection subsides, the clear ring shrinks and trade cumuli again drift in across the coast. This sequence suggests a simple orographic-convection cell, with ascending motion over the heated land and mountains, and compensating subsidence over the coastal waters. With such a picture in mind, Riehl (1947) undertook an interpretation of diurnal moisture and rainfall patterns at San Juan, Puerto Rico. His work was handicapped, however, by being confined to the use of sounding observations for the coastal region only. In Hawaii, more complicated patterns of interaction between island convection and trade winds have been studied by Leopold (1949).

Since the work of Leopold and Riehl, a theoretical study of the flow of a slightly stable air stream over a heated island has been carried out by Malkus and Stern (1953; see also Stern and Malkus 1953, Stern 1954). These theoretical results, coupled with the relatively simple shape and orography of Puerto Rico, suggested a more detailed study of the convection pattern produced by that island. Such a study, carried out on 25 June 1952, forms the subject of this paper. The basic data consist of a series of vertical soundings performed by a slow-flying aircraft on a line approximately parallel to the low-level wind. Their location relative to the island and cloud formations is mapped in Fig. 1. Figs. 2 to 5 (Plates VI to IX) depict the cloud configurations. The first sounding was made about 50 km ENE of the windward coast, in apparently undisturbed oceanic air. The second was made in the clear ring just off the windward shore; the third over the foothills where island convection was active; and the fourth near the seaward edge of the wider clear ring to the lee (WSW) of the island.

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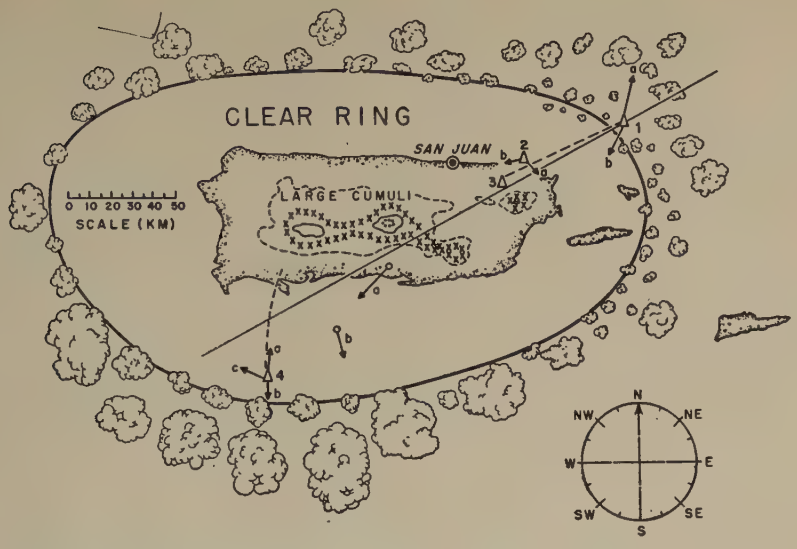


Figure 1. Map showing the topography of Puerto Rico, the locations of the four soundings, and the cloud distribution (schematically) near noon on 25 June 1952. The 1,000-ft contour is dashed; the 2,000-ft, ×-ed; the 3,000-ft solid, and the 4,000-ft, dotted. The light solid line through the sounding locations is the cross-section constructed (see Figs. 7 to 9); the arrows show the directions aimed by the camera in the photographs to follow (Figs. 2 to 5). The dashed lines running from soundings 1 and 4 to shore indicate where horizontal runs were made by the observing aircraft. The latter data are omitted from the discussion since little new information was added by them.

One of the primary reasons for performing these observations was to study the effects of divergence, subsidence, turbulence, and heating upon the various layers of the undisturbed trade-wind atmosphere, and upon the structure and height of the trade inversion. In an analysis of the Pacific trade (Riehl, Yeh, Malkus and La Seur 1951), budgets of moisture, momentum, and heat were calculated along a section of the flow about 2,500 km in length. This was a mean picture, averaged over a two months' series of observations. In it was illustrated the role of subsidence in strengthening the trade-wind inversion and the counter-working role of cumulus convection in raising and weakening it. Along the present cross-section, just over 200 km in length, the air is first subjected to subsidence (windward clear ring), then to violent turbulence, convection, and heating (over island and mountains) and then again to subsidence (leeward clear ring).

To discuss the resulting effects upon the air structure, it is first necessary to establish the motion field. This will be done after a brief presentation of the methods of measurement and several cross-sections.

2. The data

(*a*) *Method of sounding*

The basic instrumentation of the sounding aircraft was a modified form of the M.I.T. psychrograph, described previously by McCasland (1951). The plane (Catalina) was flown in helical ascents of about 2 mi diameter at air-speeds of 50-60 m/sec and at a sufficiently slow rate of climb (200 ft/min) for lag-free values of temperature and humidity to be obtained at about 30-m height intervals. The soundings are thus well suited to accurate determination of vertical gradients of these properties. A water-column accelerometer was operated to record small-scale turbulence which has been described by a parameter called ' turbulence index ' defined by Malkus and Bunker (1952). Values

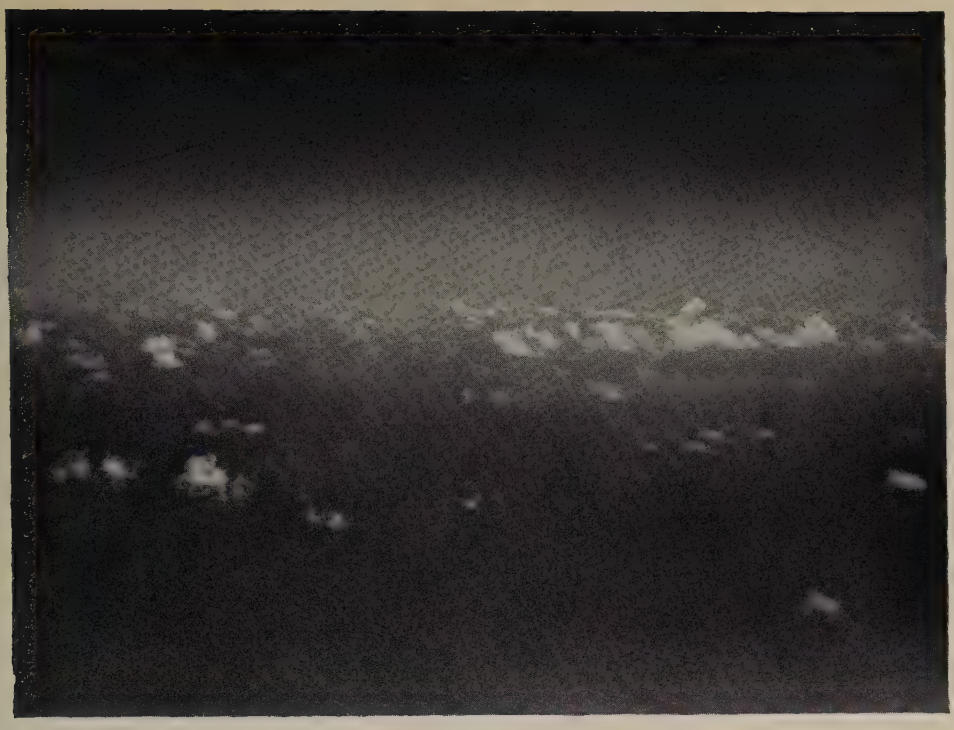
are obtained by planimetering the accelerometer trace over five-second intervals and ar only suitable for comparisons, since the actual magnitudes depend upon the aircraft Supplementary observations included photographs (Figs. 2 to 5, Plates VI to IX) an wind determinations. Surface-wind directions were ascertained by smoke drifts; ove the water, smoke flares were dropped. Winds at 1,000 ft were determined by doubl drift, using two independent sights and observers.

The four soundings are plotted in Fig. 6. Unfortunately an hour's delay occurre between the end of sounding 3 and the commencement of sounding 4 owing to the failur of the aeroplane's interphone, and sounding 4 cannot be considered synoptic with th other three. Furthermore, just as that last sounding was being made, the wind wa shifting to a few degrees south of east owing to the passage of a weak easterly wave. Thu the trajectory of the air at the location marked 4 on Fig. 1 was not along the cross-section The fact that the shift was so recent, however, and the character of sounding 4 itsel suggest that the air at that location had an over-island trajectory. Sounding 4 is thu included in the discussion, although the calculations are restricted to the remaining three which were performed in rapid succession between 1115 and 1415 LMT. Th San Juan time-section (not reproduced) shows that the weak easterly wave-system pro duced no noticeable variations in the properties of the air in the region. Upon the air craft's return to San Juan at 1730 LMT conditions there, including the height of the inversion which was checked just off shore, were the same (except for the slight wind shift) a they had been early in the morning.

(*b*) *Construction of the cross-sections*

The observations are more readily interpreted in conjunction with Figs. 7 and 8, which are sections across the island in the plane of the soundings (see map in Fig. 1). Fig. 7 shows the distribution of potential temperature and mixing ratio. Within the region delineated by the light rectangle (soundings 1-3) the picture may be considered synoptic for about noon. The remainder of the section was put together using sounding 4, theoretical calculations, other information and speculation. The potential-temperature gradient along the island surface was chosen consistent with the figures presented by Riehl (1954). He gives maximum temperatures for normal sea-breeze days at San Juan, Yayuya (elevation $\simeq$ 600 m), and Mayaguez (lee shore). Since the sea breeze on 25 June was weaker (9 mi hr^{-1}) than Riehl's median figure (15 mi hr^{-1}) his maximum gradients have been slightly reduced in Fig. 7. The mountain top in the section, for example, is 6°A potentially warmer than San Juan, while Yayuya (same elevation) often is 7-8°A potentially warmer on well-developed sea-breeze days. The potential-temperature configuration between lee shore and sounding 4 was drawn consistent with the patterns downwind of Nantucket, observed by Malkus and Bunker (1952). The potential-temperature field is everywhere consistent with the velocity field presented later. The heights to which superadiabatic and adiabatic lapse rates are shown to extend have been governed by the height of cloud base, which in general limits the top of the mixed ground layer.

Fig. 8 is a schematic reconstruction of the section. The stabilities, cloud configurations, winds at 1,000 ft, and the trade-inversion base are denoted. The latter was determined by the dual criteria of a decrease in temperature lapse-rate and a discontinuity in vertical moisture-gradient towards more rapid drying. On soundings 1, 2 and 4, these criteria were clearly met at the levels indicated. The inversion base possessed a mixing ratio of 8·7, 8·7, and 8·4 g/kg, respectively, at these locations. Between soundings 1 and 2 the additional criterion of constant potential temperature (303·4°A) could also be



(a)

(b)

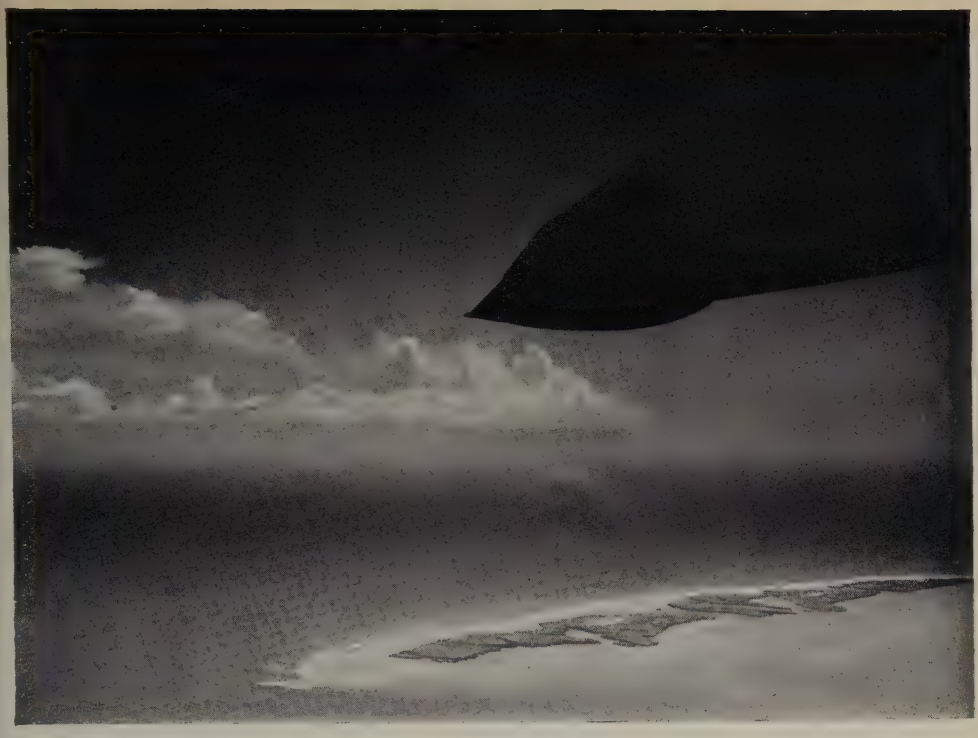
Figure 2. Photographs taken at location of sounding 1, 1200 LMT, 6,000-ft elevation. (a) Photograph of trade cumuli seaward of subsidence ring, along direction of arrow a on Fig. 1. (b) Photograph of clear ring and clouds along coastal plane (over island), along direction of arrow b on Fig. 1. Coastline is visible in right foreground.

To face page 540

(a)

(b)

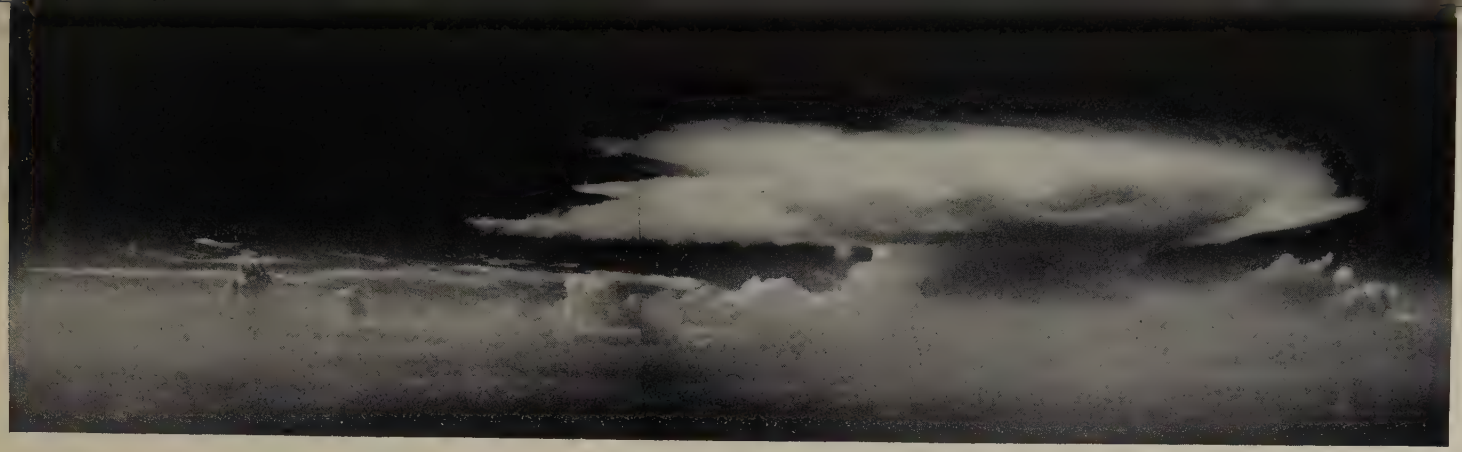
Figure 3. Photographs taken at location of sounding 2, 1333 LMT, 6,000-ft elevation. (a) Photograph along arrow a of coastal plane clouds. All clouds seen here are land-formed. (b) Photograph along arrow b of coastal plane clouds. Outskirts of San Juan are in right foreground.



(a)

(b)

Figure 4. Photographs taken on way to location of sounding 4, 1436 LMT, about 2,000-ft elevation. (a) Photograph taken at location and in direction of arrow a showing lee coast of Puerto Rico, lee side subsidence ring, and oceanic cumulus farther to seaward. (b) Photograph taken at location and in direction of arrow b showing oceanic trade cumuli to seaward of subsidence ring off lee shore.



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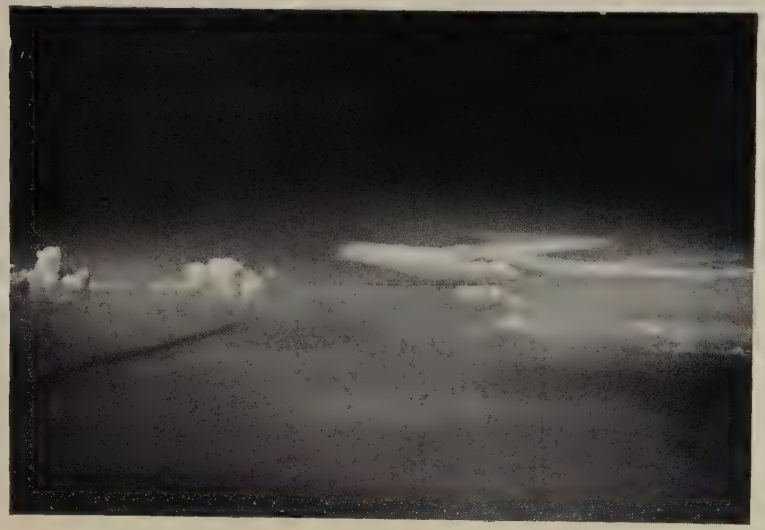

(c)

Figure 5. Photographs taken at location of sounding 4, 1600 LMT, 8,400-ft elevation. (a) Photograph along arrow a showing coastline and cumulonimbus cloud over mountain range. (b) Photograph along arrow b showing oceanic trade cumulus to south of leeward subsidence ring. (c) Photograph taken along arrow c showing land clouds to the right, clear ring, and oceanic clouds to the left.

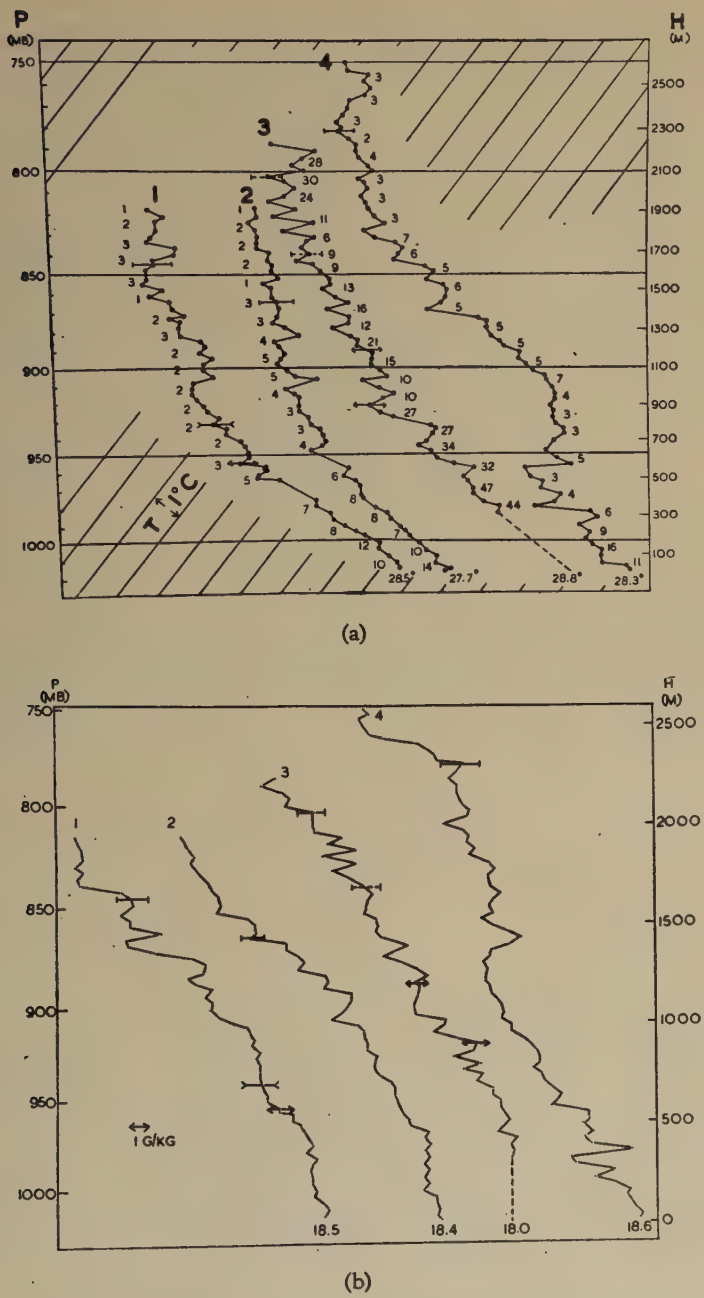

Figure 6. (a) Tephigram showing temperature (°C) as a function of pressure (left vertical axis) and height (right vertical axis) for each of the four aircraft soundings, the first sounding being on the left. The temperatures at the lowest observation point are given at the base of each sounding curve. Each curve thus has the same scale but a different origin. Dry adiabats are normal to the slanting isotherms. Sounding 3 (over land) began at 300-m elevation and was extrapolated dry-adiabatically to sea level. The figures beside each curve are values of the turbulence index averaged over 7 mb (or about 65 m) height intervals, and plotted at the centre of each interval. The trade-inversion base (not located entirely from information plotted directly on these graphs) is denoted by the lines with vertical ends; cloud base is denoted by arrows, and cloud-top level by an inverted arrow where it differed from inversion base. (b) Mixing ratio in g/kg as a function of pressure (left vertical axis) and height on a linear scale (right vertical axis) for the sounding curves of Fig. 6a. The mixing ratio at the lowest level is given below each sounding, sounding 3 being extrapolated downward at constant mixing ratio.

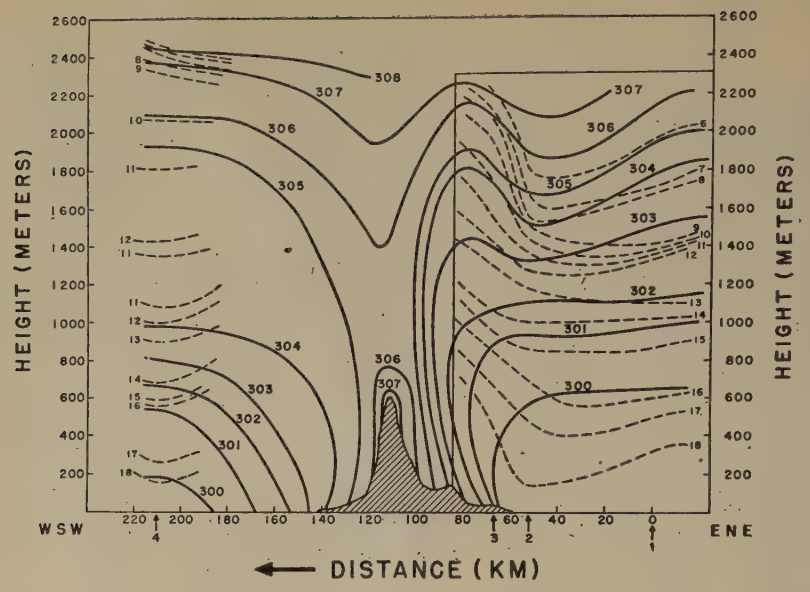

Figure 7. Cross-section of potential temperature and mixing ratio along the line indicated on Fig. 1, near noon, 25 June 1952. The contours of Puerto Rico are shown by the hatched profile. Isentropes are solid lines; lines of constant mixing ratio are dashed. The region outside the light rectangle (soundings 1 to 3) is partly speculative (see text) and not used in quantatitive calculations.

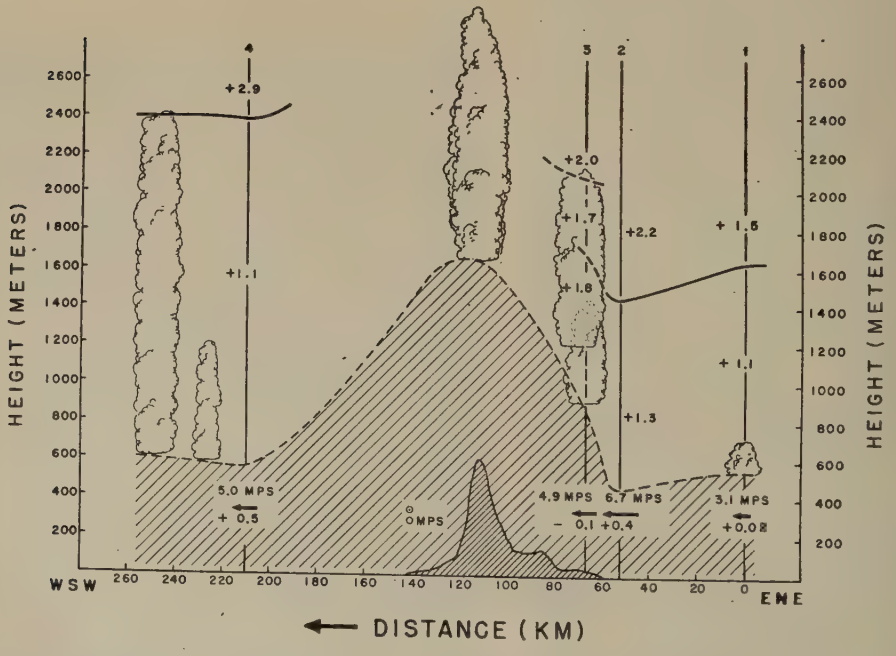

Figure 8. Schematic reconstruction of the cross-section shown in Fig. 7, giving the thickness of the subcloud layer (hatched); the distribution of the inversion base (solid, dashed on sounding 3), average stabilities for the layers indicated ($\times 10^7$ cm^{-1}), cloud configurations and 1,000-ft winds. It should be noted that the stabilities were calculated by layers and are therefore not comparable for the same height interval from one sounding to the next.

sed to follow the inversion base, since condensation and non-adiabatic heating were bsent. On sounding 3 the inversion was indefinite and appeared to have a multiple tructure, a weak transition being indicated at 1,690 m (mixing ratio 10·6 g/kg) and nother more pronounced stabilization and drying at 2,070 m (mixing ratio 8·0 g/kg). The stability is defined as $\frac{1}{\theta}\frac{d\theta}{dz}$, where θ is potential temperature. The average stability or each layer was obtained from the soundings by graphical averaging over each 100-m eight-interval and numerical averaging of these averages over the layer in question. It vas checked graphically for each layer as a whole using Fig. 6a. The heights of cloud ases and tops were measured, except for the tops over the mountains, which at noon vere estimated to be at about 10,000 ft. These built up to huge anvils surpassing 0,000 ft (see Fig. 5a) by mid-afternoon and died away near sundown.

4. The field of motion

(a) *The observed wind structure*

Surface smoke showed directions fluctuating between 60°-88° on soundings 1-3. Double drift-readings at 1,000 ft in each of these locations indicated little or no shift n direction with height. The 1100 LMT radio-wind observation at San Juan showed small wind shear up to 3 km. Thus the average 1,000-ft wind at soundings 1 and 4, namely 4 m/sec, is taken as a constant basic current throughout the section, upon which the calculated mean convective motions are to be superposed.

(b) *Calculation of the mean convective motions*

It is assumed that the sea-breeze cell is in a nearly steady state during the hours around noon, as has been justified for smaller islands by the work of Malkus and Stern (1953).

The velocities in this cell are computed as follows : a circular island of radius 50 km s considered, with an elliptical subsidence ring around it; the ring is 40 km across on he upwind side and 80 km across on the downwind side. An inward-pointing sea breeze s taken, around the coast line, of 3·6 m/sec on the upwind coast (difference between bserved winds at 1,000 ft at soundings 1 and 3), increasing to 4 m/sec on the lee coast owing to calm at 1,000 ft observed there). Inland, the sea breeze is assumed to fall linearly o zero at the island centre and, out to sea, it goes linearly to zero at the periphery of the subsidence ring. If we assume strictly radial velocities, the low-level divergence pattern s thus specified, and to complete the determination of the velocity field it is only necessary o find the height at which the sea breeze reverses. Stern and Malkus (1953) have derived an expression for this height, namely

$$h = \tfrac{1}{2}\pi (U^2/gs)^{\frac{1}{2}}$$

where U is the basic current; g is the acceleration of gravity; and s is the stability of the undisturbed current. Taking $U = 4$ m/sec, and s as the mean stability of the moist ayer on sounding 1, this height h becomes 850 m. Therefore, the horizontal-divergence ield is taken as constant from the surface to 300 m, is decreased linearly from there to zero at 900 m over the water (1,000 m over the island) and increased linearly in the opposite sense up to surface magnitude at 1,500-1,800 m. The last is not critical for the budgets which follow and was done in agreement with the theoretical results of Malkus and Stern.

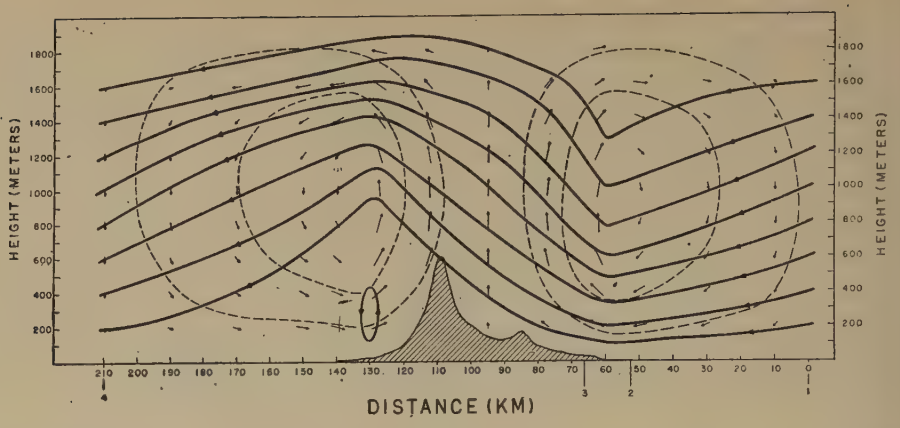

Figure 9. Streamlines (calculated) of the air flow along the cross-section of Figs. 7 and 8. The streamlin of the convection cell are dashed; the total streamlines are solid (obtained by adding constant basic wi of 4 m/sec from right to left).

Thus $\partial w/\partial z$ and w (the vertical velocity) are specified. When w is so calculate the convective and total (4 m/sec basic current added) velocities along the section giv the streamlines shown in Fig. 9. The maximum rate of descent in the subsidence rin is 3·3 cm/sec at 900 m, and the maximum average rate of ascent over the island is abou 10 cm/sec at 1,000 m. The rate of descent in the subsidence ring at 1,500 m is foun to be 1·2 cm/sec in fair agreement with the 0·8 cm/sec calculated from the sinking the inversion between soundings 1 and 2. In constructing Fig. 9, the presence of th actual mountain ridge has been ignored, for although it undoubtedly disturbs the flo and adds to the circulation, the ' equivalent mountain ' due to the heating may be show (see Stern and Malkus, Eqs. 34 and 35, 1953) to be at least $1\frac{1}{3}$ times as high as the actu 600-m mountain, provided that the average island surface-temperature exceeds that the sea by 1·5°C. Since the temperature at the lee shore (Fig. 7) is 4°C greater tha that at the windward, this mean temperature-excess is not exaggerated, and the ' equiva lent mountain ' amplitude of 800 m is consistent with the vertical amplitude of th streamlines, as also was found in the previous studies.

An independent test of the velocity field is, however, possible. The calculate motions must transport water vapour through and within the section in such a way tha mass continuity is satisfied, using the observed mixing ratios and gradients. The nex section is thus concerned with the moisture balance.

5. The moisture budget of the section

Soundings 1, 2, and 3 are assumed synoptic to describe a steady-state air flow. W consider a volume of unit width normal to this section (see Riehl *et al.* 1951) and approxi mately 2 km high. Since no precipitation was occurring in this part of the section, th processes which must achieve a balance in the lowest layer are : evaporation into th section from the surface; influx at the upstream side and efflux at the downstream side efflux or influx along the top depending on whether the mean motions are subsiding o ascending; and efflux from the top due to turbulence. Symbolically we may write

$$\rho \int_0^H U_i\, q_i\, dz + Ed - \bar{w}\, \rho\, q_H d - \rho \int_0^H U_E\, q_E\, dz - \int_0^d A \left(\frac{\partial q}{\partial z}\right)_H dx = 0$$

where the z-direction is vertically upward; the x-direction points along the basic flow

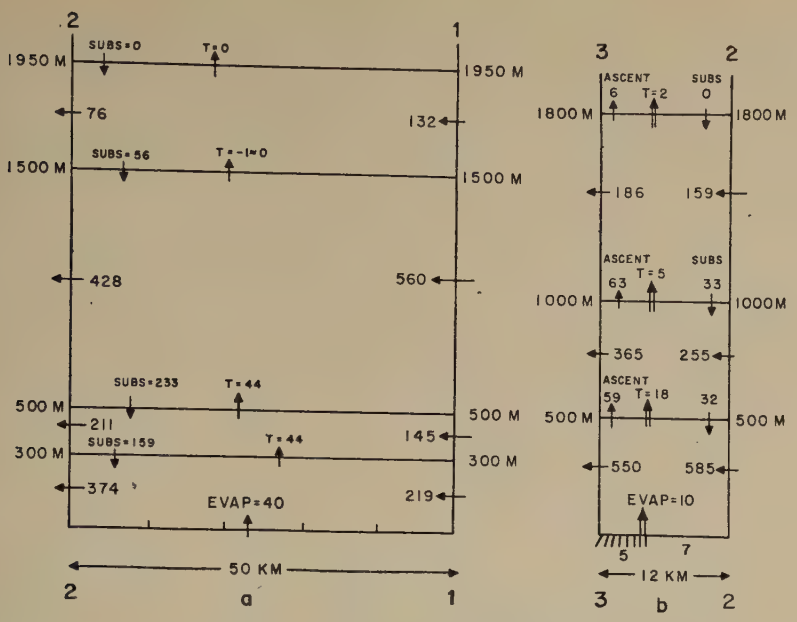

Figure 10. Moisture budget for the volume 1 cm normal to the section of Figs. 7 to 9. The numbers accompanying the arrows are moisture fluxes in g/sec into and out of the sub-volumes. (a) Budget for the volume 50 km in length between soundings 1 and 2, which was divided into four layers : lower subcloud (0-300 m); upper subcloud (300-500 m); cloud (500-1,500 m); inversion (1,500-1,950 m). The turbulent transport through the 1,500 and 1,950 m levels was zero despite a strong rate of decrease of mixing ratio upward, thus indicating a negligible Austausch at these levels. (b) Budget for the volume 12 km in length between soundings 2 and 3, which was divided into three layers : lower subcloud (0-500 m); upper subcloud (500-1,000 m); cloud (1,000-1,800 m). The much shorter length of this section makes its figures rather unreliable compared to the previous one, especially since some gradients and transports (those indicated by double arrows) were averaged across the top which includes a water-land transition. Nevertheless the results are consistent with one another and with the sounding data.

the section has a height H and a length d. The first term is the influx at the upstream end due to the horizontal motion along the section, U_i; q_i is the mixing ratio at the influx end; and ρ is the air density, assumed constant throughout the section. The second term is the moisture addition due to evaporation, E being the evaporation rate per unit area. The third term is the flow through the top of the section, where $\bar{w}$ is the mean vertical motion along the length of the top, and q_H is the mean mixing-ratio along the top. The fourth term is the horizontal efflux at the downstream end and the last term is the efflux through the top due to turbulence. The vertical gradient of moisture at the level H is $\partial q/\partial z$ and A is the Austausch or coefficient of turbulent moisture exchange. All these quantities except the evaporation rate and the Austausch are determined by the field of motion and the soundings. If the evaporation rate can be estimated, the Austausch at the upper boundary of the lowest layer may be calculated and a similar budget-equation evolved for the next higher layer of the section. Since two terms are unknown for the first sub-section, it might appear that quite a wide leeway is permitted in their choice and that the field of motion is not thereby rigorously tested. The calculations show, however, that small variations in the velocity field require rather implausible choices of evaporation and Austausch distribution to achieve a balance in all the subsections, since the fluxes in each one depend upon those in all the others, both vertically and in the horizontal (see Fig. 10). Large variations in velocity field prohibit balance in one or more sub-sections under any reasonable assumptions concerning the remaining two fluxes.

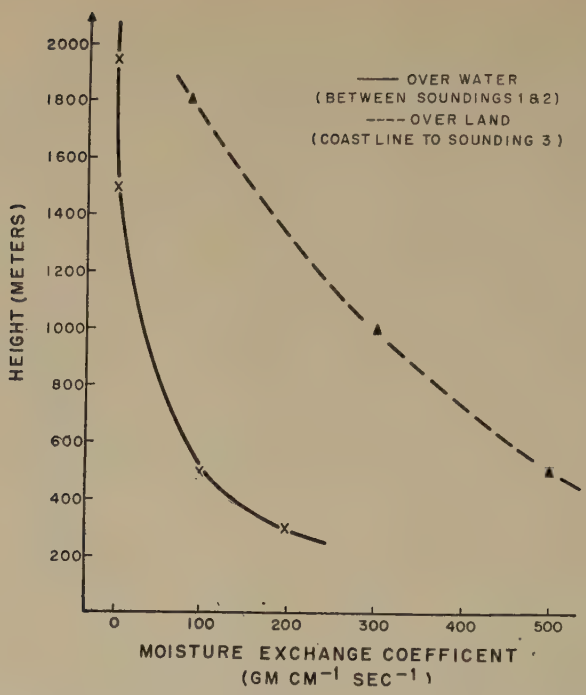

Figure 11. Austausch, or moisture-exchange coefficient, as a function of height, calculated from the result of Fig. 10 taking the turbulent transport, T, and the vertical moisture gradient at the level in question. The values over land (dashed curve) are computed by obtaining the average value between soundings 2 and 3 (Fig. 10b) and extrapolating the value over water (solid curve) up to the coastline.

An evaporation rate has therefore been taken which is plausible and gives a reasonable distribution of Austausch. The process is illustrated in detail for the lowest 300-m layer between soundings 1 and 2, and the remaining calculations are shown in Figs. 10 and 11. The Austausch distribution over land was found from the average between soundings 2 and 3 by extrapolating the overwater value up to the coastline.

The figures for the lowest 300-m section between soundings 1 and 2 are as follows:

$$\text{Horizontal influx term} = \rho \overline{U}_i \bar{q}_i H = 10^{-3} \times 4 \times 10^2 \times 18{\cdot}2 \times 10^{-3} \times 3 \times 10^4 = 219 \text{ g/sec}$$

$$\text{Horizontal efflux} = \rho \overline{U}_E \bar{q}_E H = 10^{-3} \times 7 \times 10^2 \times 17{\cdot}8 \times 10^{-3} \times 3 \times 10^4 = 347 \text{ g/sec}$$

$$\text{Subsidence through top} = \rho \overline{w} q_{300} d = 10^{-3} \times 1{\cdot}8 \times 17{\cdot}6 \times 10^{-3} \times 50 \times 10^5 = 159 \text{ g/sec}$$

These figures are fixed by the soundings and calculated velocities. An evaporation rate of 8×10^{-6} g cm^{-2} sec^{-1} is now assumed, which is twice the mean annual evaporation rate for the region (Sverdrup 1942). This gives the remaining two fluxes as:

$$\text{Evaporation} = Ed = 8 \times 10^{-6} \times 50 \times 10^5 = 40 \text{ g/sec}$$

$$\text{Turbulent flux} = A_{300} (\partial q/\partial z)_{300} d = A_{300} \times 5 \times 10^{-9} \times 50 \times 10^5$$

Applying the balance equation, A_{300} is found to be 176 g cm^{-1} sec^{-1}, or 200 g cm^{-1} sec^{-1} within the errors of the calculations. Clearly the balance within this sub-section can be achieved with many alternative arrangements between the last two fluxes, including equating both of them to nearly zero. However, when the remaining three upper and three downstream sections are considered it will be seen that those two fluxes can be varied only little if plausible balances are required throughout, since the fluxes into each section depend on the effluxes from the neighbouring ones.

The height distribution of Austausch coefficient in Fig. 11 compares well with values obtained for the Caribbean (Bunker, Haurwitz, Malkus and Stommel 1949) and in the Pacific trade by Riehl *et al.* (1951). The higher values and greater vertical extent of the turbulence over land are also indicated by the much higher ' turbulence index ' on sounding 3 compared with the others (see Fig. 6a).

5. Effects upon the tropical air mass on passing through the convection cell

It is now possible to return to the cross-sections to examine the effects of the processes established above upon the meteorologically important features of the trade stream. For this purpose it is again convenient to divide the section into vertically superposed layers.

(*a*) *The properties of the subcloud or ground layer*

The subcloud layer is defined in Fig. 8 as the layer of air between the surface and the region of slight stabilization and drying which usually marks the level of cloud base. The subcloud layer exhibited by sounding 1 is typical of undisturbed oceanic trade-wind air (see Bunker *et al.* 1949). Its depth is 575 m; it is barely stable with respect to dry adiabatic processes; it is also well-mixed and homogeneous with a nearly constant mixing-ratio. At the location of sounding 2, the character of the layer is markedly changed. Subsidence and divergence have caused it to shrink to 450 m depth; *it has stabilized by a factor of about twenty*, and dried slightly (mean mixing ratio reduced by 5 per cent compared to sounding 1).

Over the plains, the lapse rate is superadiabatic up to 300 m (using the 1100 LMT San Juan sounding up to the height where the aeroplane sounding commenced) and adiabatic from there up to 900 m, the lowest height of a highly variable cloud base. The mean stability in the lowest 900-m layer is negative and the moisture uniformly distributed. Turbulence is intense and the turbulence index reached extreme values. On the last sounding, at the outer edge of the leeward subsidence ring, the subcloud layer has shrunk again to about 575-m thickness, but its structure is markedly different from the normal oceanic ' homogeneous ' layer. The mean stability is even greater than that at sounding 2 and the average moisture lapse-rate is three times greater than in the undisturbed trade-wind air at sounding 1. Most striking, however, are the intense fluctuations in mixing ratio with height, indicating that the layer was not well mixed at all, but suggesting that masses of drier air had been brought down from aloft and not yet incorporated.

(*b*) *The relation between the structure of the subcloud layer and the initiation of cumulus growth*

Small trade cumuli uniformly based at 575 m are present at sounding 1, while clouds are entirely absent at sounding 2. Since calculations using normal rates of mixing show that the lowest 300 m of the ' cloud ' layer at sounding 2 are less inhibitory (by drying) to cumulus growth than at sounding 1, the reason for the suppression of cumuli at sounding 2 must be sought in the subcloud layer. The work of Bunker *et al.* (1949) indicated that trade cumuli generally were found in localities where the lifting condensation level (LCL) of the lowest air fell below the top of the mixed layer, and clear spaces occurred where the LCL was more than about 30-50 m above its top. At sounding 1, the LCL of the lowest air

is 585 m, the same within observational error as the top of the mixed layer and the observed cloud base. At sounding 2, the LCL of the lowest air is 560 m, which exceeds the top of the mixed layer by 110 m. Sounding 4, however, suggests that a lifting condensation level close to the top of the ground layer is not a sufficient condition for cumulus formation. No clouds appeared at sounding 4, although the LCL for the lowest air was at 600 m. Apparently the stability of the subcloud layer inhibited the upward mixing of water vapour from the sea surface and the turbulent eddying necessary for the initiation of clouds. A few kilometres seaward of sounding 4, small cumuli based at 600 m began to appear, increasing in size still farther off shore, suggesting that the mixed layer and its transport functions were re-established after leaving the region influenced by subsidence.

At sounding 3 over the plains of Puerto Rico, cloud-base height ranged from 900 to about 1,200 m. Here the heating condensation level is probably more pertinent than the LCL. For air with a mean mixing ratio of 1 g/kg (or 6 per cent) in excess of that sounding, the HCL was at 900 m and no further heating was required for cloud production since the lapse rate is adiabatic up to that level. For air with the mixing ratio of that sounding, the HCL was at 1,275 m, requiring a 1°C increase in surface temperature. Local fluctuations in temperature and moisture can therefore easily account for the height variations in cloud base. In contrast to the oceanic cumuli, it appears that the land clouds were associated with thermals extending well below their bases, since cloud-scale 'bumps' could be traced under clouds down to the lowest level flown (approximately 300 m).

(c) *The properties of the cloud layer*

The cloud layer is defined as lying between the top of the ground layer and the base of the trade inversion, regardless of whether clouds may actually be present. One obvious variation in that layer was the great rise in cloud base from 575 m over water, to 900-1,200 m over the plains, and to more than 1,600 m in the mountainous area. This common island effect is explained by more rapid warming than moistening of the lower air. A second is the very different appearance of the clouds themselves (see Figs. 2 to 5). The land and mountain clouds exhibit an appearance similar to middle-latitude cumulus, with hard, cauliflowerlike edges and boiling motions, in contrast to the softer, more slanting, more slowly overturning trade cumuli. The stability in the cloud layer is, however, far greater over the land than over the ocean, being nearly the same at soundings 1, 2, and 4, and larger by 50 per cent at sounding 3. This feature (namely stabilization over land in the middle and upper cloud layer) was noted by Riehl (1947) who attributed its cause to the intense turbulence below, which creates a stable region at its upper limit. Many of the clouds over the plains terminated at the first weak inversion at 1,690 m and the tallest ones stopped at the stronger one at 2,070 m.

Downstream variations in cloud layer humidity are also pertinent. Between soundings 1 and 2, the average mixing ratio of the cloud layer (roughly 500-1,500 m vertical extent) decreased by 13 per cent. On soundings 3 and 4 the cloud layer has thickened so much that mean mixing ratios are not comparable. The difference in mixing ratio between sounding air and saturated air at the same temperature was therefore taken as a measure of the probable difference in vapour content between clouds and environment. For the 1,500-m level, at sounding 1 this difference is 6·2 g/kg; at sounding 2, it is 6·0 g/kg, at sounding 3, 3·3 g/kg; and at sounding 4, 4·4 g/kg. Thus, over the island, the greater stability in the cloud layer was far more than offset by three factors, (a) strong thermals below cloud base, (b) general ascent in the mean motion, and (c) reduced resistance due to entrainment (a less dry environment).

(d) *Features of the moist layer as a whole and of the trade-wind inversion*

In the tropics, the moist layer is defined as extending from the surface up to inversion base. Sounding 4 shows clearly the effects of passage across the island; the moist layer extends up to 2,300 m, compared to 1,600 m on sounding 1. Relative to sounding 1, the air from 500-1,300 m has dried, decreasing its mean mixing ratio by 10 per cent, and the air from 1,300-2,000 m has been moistened increasing its mean mixing ratio by 38 per cent. Fig. 9 shows that the effects of the subsidence and divergence had become small by 1,300 m, while mean ascent and cumulus activity over the land reached far greater heights. The net effects of passage through the convection cell were thus such that the moist layer as a whole accumulated water vapour.

We saw that the trade inversion-layer sank 200 m between soundings 1 and 2, stabilized, and conserved its potential temperature. After the coastline is passed, the isentropes begin to cross the inversion contours very rapidly and at sounding 4 the potential temperature of the base of the layer has increased by 4·5°C. Level for level, sounding 4 shows (compared with sounding 1) a net warming up to 1,000 m, a net cooling above about 2,000 m, and a marked destabilization between, to be expected as a result of the convective overturning. The intensified inversion on sounding 4, both in respect of stability and vertical rate of drying, must be due to the fact that the inversion marks the upper boundary of the convection, rather than to the subsidence, which vanished well below that level. Thus although it may be useful throughout such a section to refer to the stable boundary between moist and dry air as ' the ' trade inversion, this inversion is physically continuous only between soundings 1 and 2. Thereafter it is controlled in strength and height by differing processes at different stages as the air passes through the cell.

6. Conclusions

In addition to establishing the existence and magnitude of the orographic-convection cell suggested visually, the present study illustrates the critically sensitive balance between the processes normally at work in the trade stream. An island serves to intensify some of these processes relative to others as the air passes across it and thus suddenly alters the balance, quickly bringing about results which may occur imperceptibly slowly on a long over-ocean trajectory.

Particular attention has been focused on the moisture budget and upward transport of water vapour. The trade cumuli are the main effectors of this transport through a deep layer, and their suppression over large areas or times may cause not only local drought but a disturbing influence on larger-scale circulations. It has been shown here that small alterations in the subcloud layer due to subsidence may prevent the appearance of cumuli, despite little or no change in the cloud layer itself. This is due to the marginal condition of the normal subcloud layer. Under ordinary conditions, a decrease in lapse rate of less than 0·1°C per 100 m may increase the stability by more than a factor of five; or a decrease of mixing ratio of 0·5 g/kg at the top of the layer may double the moisture gradient. Either of these changes alters the eddy transports through the layer. Subsidence is a particularly effective suppressor of cumuli since it both stabilizes and thus damps the eddies and may also lower the top of the mixed layer below the lifting condensation level of the air within it. At higher levels, subsidence serves to lower and strengthen the trade inversion, the contours of which we see from Fig. 8 would be far different without the normally observed counteraction of the cumuli.

A heated land mass produces the opposite effects. Over an island the size of Puert Rico on even a slightly subnormal sea-breeze day, the height of cloud top is raised fro an undisturbed 700 m over water to 3 km over land by noon and to more than 9 km b mid-afternoon. After leaving the island the air had experienced a net 800-m raising c the trade inversion and a 23 per cent accumulation of water vapour in the moist laye Clearly an observer's picture of the trade-wind atmosphere will depend critically upo whether he observes over water or over land, and, if over land, upon the size of the lan mass and his particular location with respect to it.

Finally the very delicate equilibrium governing the trade-wind air stream raises th possibility of artificial attempts to alter it in limited areas. In particular, the critic condition of the subcloud layer indicates that man-made efforts to increase tropic rainfall might more fruitfully be concentrated upon the air below cloud and upwind c the islands rather than upon the already existent clouds over the region where rain i desired, which this study suggests are in a far less marginal balance.

Acknowledgments

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Theory of non-horizontal adiabatic flow in the atmosphere

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Summary

The equation of continuity takes the form

$$\text{div } \mathbf{v} = gw/c^2$$

when there are appreciable vertical displacements (see Section 1 for list of symbols). The stream function, ψ, for two-dimensional steady motion satisfies an equation which for vertical displacements of less than about a kilometre reduces to the form

$$\nabla^2 \psi = G_1(\psi) - z\, G_2(\psi),$$

where G_1 and G_2 are functions of ψ determined by the profile of the air current entering the region under study. This is probably the best form for numerical solution.

Bernoulli's equation shows that $(gz + q^2/2 - \varpi/\tau)$ is constant along a streamline in steady adiabatic flow.

The perturbation equations which have already been widely used are derived as a special case. Except in this case the solution for three-dimensional flow is very much more difficult to obtain than that for two-dimensional flow.

The lateral stretching of vortex lines in flow over unlevel ground reduces or increases the surface wind according as the wind increases or decreases with height. This process may be used to explain the location of blow-downs in forests on hillsides.

The generation of shear in non-horizontal motion can occur at already existing discontinuities of temperature (or velocity). If the motion is unstable so that turbulence is generated, other things being equal the more stable (statically) the temperature inversion the more likely it is to produce turbulence in mountain waves.

1. Introduction

Considerable success has been achieved in understanding the flow of air over mountains by means of the solution of the equations of motion in linearized form (Scorer 1949-54). The success has in some respects been surprising and it therefore seemed desirable to obtain an equation for the flow in which it is not necessary to assume that the disturbance from horizontal flow is infinitesimal. Long (1953) has given an equation for the flow of a stably-stratified incompressible fluid and has applied it to the interpretation of model experiments. The main purpose of the present paper is to derive a similar equation for the atmosphere, which is compressible, and to see in what respects it is similar to the incompressible case and in what respects perturbation theory is satisfactory.

Many approximations are necessary; they are not made arbitrarily as in the linearized equations, but according to the magnitude of the terms in the kind of flow under discussion. The main difference due to compressibility is in the equation of continuity : this arises because the air expands or contracts when subjected to vertical displacements.

List of symbols used

c velocity of sound
$\mathbf{f}$ twice the earth's angular velocity
$\mathbf{g}$ gravity $(0, 0, -g)$
k wave number
l defined by Eq. (7·6)
p pressure, p_1 = standard value, e.g. 1,000 mb

q scalar velocity
R gas constant
t time
T temperature
$\mathbf{v}$ velocity vector (u, v, w)
U, V horizontal velocity components in undisturbed stream
x, y, z Cartesian coordinates, z vertical
β static stability defined by Eq. (2·4); β_0 = value in undisturbed stream
γ ratio of specific heats of air
ϵ $g\zeta/c^2$
ζ vertical displacement of particle from original level
θ potential temperature
ϖ modified pressure defined by (2·1)
ρ density, ρ_0 = value in undisturbed stream
τ $1/\theta$
ψ stream function defined by Eq. (4·1)
A prime denotes $\partial/\partial z_0$, z_0 being the level from which the particle at a point originated.

2. The 'modified pressure'

The modified pressure ϖ is related to the pressure by the equation

$$\varpi = \frac{\gamma R}{\gamma - 1}\left(\frac{p}{p_1}\right)^{\frac{\gamma-1}{\gamma}} \quad . \quad . \quad . \quad . \quad . \quad (2.1)$$

The potential temperature θ is defined by

$$\theta = \frac{1}{\tau} = T\left(\frac{p_1}{p}\right)^{\frac{\gamma-1}{\gamma}} \quad . \quad . \quad . \quad . \quad . \quad (2.2)$$

and so

$$\frac{1}{\rho}\operatorname{grad} p = RT \operatorname{grad}\log p = \frac{\gamma R}{\gamma - 1} T \operatorname{grad}\log \varpi$$
$$= \frac{1}{\tau}\operatorname{grad}\varpi \quad . \quad . \quad . \quad . \quad . \quad . \quad (2.3)$$

The static stability is defined as

$$\beta = \frac{1}{\theta}\frac{\partial\theta}{\partial z} = -\frac{1}{\tau}\frac{\partial\tau}{\partial z} \quad . \quad . \quad . \quad . \quad . \quad (2.4)$$

3. The equation of continuity for adiabatic motion in the atmosphere

In adiabatic motion

$$\frac{Dp}{Dt} = c^2\frac{D\rho}{Dt} \quad . \quad . \quad . \quad . \quad . \quad (3.1)$$

The equation of continuity is

$$\frac{D\rho}{Dt} = -\rho \operatorname{div}\mathbf{v} \quad . \quad . \quad . \quad . \quad . \quad (3.2)$$

and the equation of inviscid motion in the atmosphere is

$$\frac{D\mathbf{v}}{Dt} = -\frac{1}{\rho}\operatorname{grad} p + \mathbf{g} + \mathbf{v}_A\mathbf{f} \quad . \quad . \quad . \quad (3.3)$$

) that taking the scalar product with **v** we have

$$\frac{Dp}{Dt} = \frac{\partial p}{\partial t} + (\mathbf{v} \,.\, \text{grad})\, p$$

$$= \frac{\partial p}{\partial t} - \rho \mathbf{v} \,.\, \frac{D\mathbf{v}}{Dt} + \rho \mathbf{v} \,.\, \mathbf{g}$$

$$= \frac{\partial p}{\partial t} - \rho \left\{ \frac{D}{Dt} \frac{q^2}{2} + gw \right\} \quad . \quad . \quad . \quad . \quad (3.4)$$

Combining this with (3·1) and (3·2) we obtain

$$\text{div } \mathbf{v} = -\frac{1}{\rho c^2} \frac{\partial p}{\partial t} + \frac{1}{c^2} \left\{ \frac{D}{Dt} \frac{q^2}{2} + gw \right\} \quad . \quad . \quad . \quad (3.5)$$

The terms on the right-hand side of this equation represent the divergence due to changes n pressure, and therefore contain the velocity of sound for this is one of the quantities by hich we can represent the elastic properties of the air. Only if the motion is very un-teady is the term $\rho^{-1} \partial p/\partial t$ comparable with the term gw : for instance the two terms re about equal if the local rate of change of pressure is 1 mb/min and $w = 15$ cm/sec. The vertical velocities accompanying such rapid changes of pressure are usually many mes this amount. Furthermore we are primarily interested in steady motion because ne motions accompanying such rapid pressure changes as 1 mb/min are either intractable y the present methods (e.g. the case of violent convection), or they can be treated as roblems in steady motion by translation of the axes with the waves, or they accompany ompression waves which are outside the scope of this paper. If the waves are not com-ression waves the Mach number, q/c, is small and so the second term is small compared ith div **v**. The term gw/c^2 represents the compression or expansion due to vertical notion in the hydrostatic pressure field. The displacements we shall be concerned with roduce changes in pressure of air parcels greatly in excess of local departures from the ydrostatic field (the accelerations in the motion are small compared with g) and so only ne third term of the right-hand side of Eq. (3.5) is retained and the equation of continuity or adiabatic motion is

$$\text{div } \mathbf{v} = gw/c^2 \quad . \quad . \quad . \quad . \quad (3.6)$$

We may note, incidentally, that the earth's rotation (included in Eq. (3.3)) does not ffect this result. It applies to three-dimensional and to unsteady motion in the atmosphere nd is therefore more general than is required for present purposes.

4. The stream function, in two-dimensional flow

Suffix 0 denotes values of quantities at some ' starting point ' on each streamline. n the two-dimensional motion here considered the starting point is thought of as being n the plane $x = x_0$ where the motion is assumed to be horizontal. Each streamline is herefore associated with a set of values z_0, θ_0, and U (the velocity, parallel to the x-axis, $x = x_0$). Then ψ is defined by

$$\partial \psi / \partial z_0 = -U \quad . \quad . \quad . \quad . \quad . \quad (4.1)$$

n adiabatic motion potential temperature is invariant and so θ and τ may be treated as nctions of z_0 or of ψ only.

We saw, in Section 3, that the greatest pressure changes are produced by vertic displacements. Thus if ζ, i.e. $(z - z_0)$, is the vertical displacement, we have approx mately

$$T = T_0 - \frac{(\gamma - 1)\, g}{\gamma R}\, \zeta \quad . \quad . \quad . \quad . \quad . \quad (4.$$

if the motion is adiabatic ($(\gamma - 1)\, g/\gamma R$ is the adiabatic lapse rate).

Continuity along a stream tube in two-dimensional flow requires, for adiabatic motio that

$$\begin{aligned} \frac{\partial z_0}{\partial z} &= \frac{\rho u}{\rho_0 U} = \frac{u}{U}\left(\frac{T}{T_0}\right)^{\frac{1}{\gamma-1}} \\ &= \frac{u}{U}\left(1 - \frac{(\gamma-1)\, g}{c_0^2}\, \zeta\right)^{\frac{1}{\gamma-1}} \\ &= (1 - \epsilon)\, u/U \quad . \quad . \quad . \quad . \quad . \quad (4. \end{aligned}$$

Strictly, ϵ is defined by this equation. It is approximately equal to $g\zeta/c_0^2$ which is abo 1/10 for displacements of the order of 1 km. Since we find that it appears only in t less important terms in the equations to be derived we shall write

$$\epsilon = \gamma\zeta/c_0^2 \quad . \quad . \quad . \quad . \quad . \quad . \quad (4.$$

and neglect terms in ϵ^2 whenever they appear.

On this basis we obtain the following results which are required later :

$$\left.\begin{aligned} \frac{\partial \tau}{\partial z} &= \frac{d\tau}{dz_0}\frac{\partial z_0}{\partial z} = -\frac{\beta_0 \tau}{U}(1 - \epsilon)\, u \\ \text{and} \quad \frac{\partial \tau}{\partial x} &= \frac{d\tau}{dz_0}\frac{\partial z_0}{\partial x} = -\beta_0 \tau\left(-\frac{w}{u}\right)\frac{\partial z_0}{\partial z} \\ &= \frac{\beta_0 \tau}{U}(1 - \epsilon)\, w \end{aligned}\right\} \quad . \quad . \quad (4.$$

$$\left.\begin{aligned} u &= (1 + \epsilon)\, U\frac{\partial z_0}{\partial z} = -(1 + \epsilon)\frac{\partial \psi}{\partial z} \\ \text{and} \quad w &= -u\frac{\partial z_0}{\partial x}\Big/\frac{\partial z_0}{\partial z} = -u\frac{\partial \psi}{\partial x}\Big/\frac{\partial \psi}{\partial z} \\ &= (1 + \epsilon)\frac{\partial \psi}{\partial x} \end{aligned}\right\} \quad . \quad . \quad (4.$$

$$\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} = -(1 + \epsilon)\, \nabla^2 \psi - \left(\frac{\partial \psi}{\partial z}\frac{\partial \epsilon}{\partial z} + \frac{\partial \psi}{\partial x}\frac{\partial \epsilon}{\partial x}\right) \quad . \quad . \quad (4.$$

$$u^2 + w^2 = (1 + 2\epsilon)\, (\nabla\psi)^2 \quad . \quad . \quad . \quad . \quad . \quad (4.$$

Also $$\frac{\partial \psi}{\partial x} = -U\frac{\partial z_0}{\partial x}, \qquad \frac{\partial \psi}{\partial z} = -U\frac{\partial z_0}{\partial z}$$

and so

$$\left.\begin{aligned} \nabla^2 \psi &= -U\nabla^2 z_0 - U'(\nabla z_0)^2 \\ (\nabla\psi)^2 &= U^2 (\nabla z_0)^2 \end{aligned}\right\} \quad . \quad . \quad (4.$$

where $U' = dU/dz_0$ which is the shear far upstream where the motion is horizontal.

Since $z_0 = z - \zeta$ we have

$$\left.\begin{aligned} \nabla^2 z_0 &= -\nabla^2 \zeta \\ (\nabla z_0)^2 &= (\nabla\zeta)^2 - 2\frac{\partial\zeta}{\partial z} + 1 \end{aligned}\right\} \quad . \quad . \quad (4.10)$$

The equation of continuity (3.6) is

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} - \frac{D\epsilon}{Dt} = 0 \quad . \quad . \quad . \quad . \quad . \quad (4.11)$$

In this equation, and in equations involving ϵ, we do not distinguish between c^2 and c_0^2 because the terms concerned are small. Thus in Eq. (4.7) the last term is

$$\begin{aligned} \frac{\partial\psi}{\partial z}\frac{\partial\epsilon}{\partial z} + \frac{\partial\psi}{\partial x}\frac{\partial\epsilon}{\partial x} &= -U\frac{g}{c^2}\left(\frac{\partial z_0}{\partial z}\frac{\partial\zeta}{\partial z} + \frac{\partial z_0}{\partial x}\frac{\partial\zeta}{\partial x}\right) \\ &= -U\frac{g}{c^2}\left(\frac{\partial\zeta}{\partial z} - (\nabla\zeta)^2\right) \quad . \quad . \quad . \quad (4.12) \end{aligned}$$

5. The equation for disturbed motion

By (2.3) the equations of motion are

$$\frac{Du}{Dt} = -\frac{1}{\tau}\frac{\partial\varpi}{\partial x}; \qquad \frac{Dw}{Dt} = -\frac{1}{\tau}\frac{\partial\varpi}{\partial z} - g \quad . \quad . \quad . \quad (5.1)$$

Eliminating ϖ we have

$$\frac{Du}{Dt}\frac{\partial\tau}{\partial z} - \frac{Dw}{Dt}\frac{\partial\tau}{\partial x} - g\frac{\partial\tau}{\partial x} + \tau\left(\frac{\partial}{\partial z}\frac{Du}{Dt} - \frac{\partial}{\partial x}\frac{Dw}{Dt}\right) = 0 \quad . \quad . \quad (5.2)$$

By means of Eqs. (4.5), and noting that $\frac{D}{Dt}\frac{\beta_0}{U} = 0$ and that $w = \frac{D\zeta}{Dt}$, this leads to

$$\begin{aligned} &-(1-\epsilon)\frac{D}{Dt}\left\{\frac{\beta_0}{U}\frac{u^2+w^2}{2} + \frac{g\beta_0}{U}\zeta\right\} \\ &\quad + \frac{D}{Dt}\left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\right) + \left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\right)\left(\frac{\partial u}{\partial z} + \frac{\partial w}{\partial x}\right) = 0 \quad . \quad . \quad (5.3) \end{aligned}$$

Using Eqs. (4.6) to (4.8) to write this in terms of ψ we obtain, neglecting terms in ϵ^2,

$$\begin{aligned} &(1-\epsilon)\frac{D}{Dt}\left\{\frac{\beta_0}{2U}(1+2\epsilon)(\nabla\psi)^2 + \frac{g\beta_0}{U}\zeta\right\} \\ &\quad + \frac{D}{Dt}\left\{(1+\epsilon)\nabla^2\psi + \frac{\partial\psi}{\partial z}\frac{\partial\epsilon}{\partial z} + \frac{\partial\psi}{\partial x}\frac{\partial\epsilon}{\partial x}\right\} + \nabla^2\psi\frac{D\epsilon}{Dt} = 0 \quad . \quad (5.4) \end{aligned}$$

On rearrangement this becomes

$$\begin{aligned} &(1-\epsilon)\frac{D}{Dt}\left\{\nabla^2\psi + \frac{\beta_0}{2U}(\nabla\psi)^2 + \frac{g\beta_0}{U}\zeta\right\} \\ &\quad + \frac{D}{Dt}\left\{2\epsilon\nabla^2\psi + \frac{\epsilon\beta_0}{U}(\nabla\psi)^2 + \frac{\partial\psi}{\partial z}\frac{\partial\epsilon}{\partial z} + \frac{\partial\psi}{\partial x}\frac{\partial\epsilon}{\partial x}\right\} = 0 \quad . \quad . \quad (5.5) \end{aligned}$$

Since the second term is of the order of ϵ, so also is the first, therefore the ϵ in its coefficient

may be dropped. The equation may then be integrated to give, using Eq. (4.12),

$$(1+2\epsilon)\left\{\nabla^2\psi+\frac{\beta_0}{2U}(\nabla\psi)^2+\frac{g\beta_0}{U}\zeta\right\}$$
$$-2\epsilon\frac{g\beta_0}{U}\zeta-\frac{Ug}{c^2}\left\{\frac{\partial\zeta}{\partial z}-(\nabla\zeta)^2\right\}=\text{func}(\psi) \quad . \quad . \quad . \quad (5.6)$$

Again we may drop the 2ϵ in the coefficient of the first term. The function of ψ (or of z) is determined by the value of the left-hand side where U and β_0 are defined.

By means of Eqs. (4.9) and (4.10) this may be converted into an equation for ζ, namely

$$\nabla^2\zeta+\left(\frac{\beta_0}{2}+\frac{g}{c^2}-\frac{U'}{U}\right)(\nabla\zeta)^2$$
$$+\left(\frac{2U'}{U}-\beta_0-\frac{g}{c^2}\right)\frac{\partial\zeta}{\partial z}-\frac{g\beta_0}{U^2}\zeta\left(1+\frac{2g}{c^2}\zeta\right)=0 \quad . \quad . \quad (5.7)$$

First we notice that the term $\frac{\beta_0}{2U}(\nabla\psi)^2$ has led to the term $\frac{\beta_0}{2}(\nabla\zeta)^2$. In typical cases $\beta\simeq 2\times10^{-7}\,\text{cm}^{-1}$ and $g/c^2\simeq 10^{-6}\,\text{cm}^{-1}$ so that the term in question is small compared with the term $(g/c^2)(\nabla\zeta)^2$ which is one of the ϵ terms. Thus as a first approximation if the terms in ϵ are omitted, Eq. (5.6) is simply

$$\nabla^2\psi+\frac{g\beta_0}{U}\zeta=U' \quad . \quad . \quad . \quad . \quad . \quad (5.8)$$

which is probably the simplest form for numerical solution, if such were attempted. It is Long's equation without the term in $(\nabla\psi)^2$ which, we now see, is negligible in the meteorological case. U' is the shear where $\zeta=0$, where also w is assumed to be zero. Since $\zeta=z-z_0=z+\int\frac{d\psi}{U}$, U being a function of ψ only, Eq. (5.8) takes the form

$$\nabla^2\psi=G_1(\psi)-zG_2(\psi) \quad . \quad . \quad . \quad . \quad (5.9)$$

G_1 and G_2 being given functions of ψ which are determined by the nature of the airstream where it flows into, or out of, the region under study. It is outside the scope of the present paper to discuss the boundary conditions which must be applied to this non-linear equation in order to obtain solutions for an infinitely deep atmosphere.

If $U'/U\simeq g/c^2$ the term in $(\nabla\zeta)^2$ in Eq. (5.7) disappears, for practical purposes. This is quite a common case, particularly on occasions when waves have been observed, although it is clearly not true if the wind decreases with height. It represents, for instance, a doubling of the wind speed between heights of 5 km and 10 km. The only non-linear term remaining is then the last. We have seen already that $(2g/c^2)\zeta\simeq\frac{1}{5}$ for a vertical displacement of 1 km. In such a height change the quantity $g\beta_0/U^2$ changes by as much as half its value in the airstreams known to produce the most interesting flows, namely those with large vertical variations in stability and wind strength. This means that no important physical effect is omitted from the equations if the term in ζ^2 is omitted. The coefficients are, even then, not constants : indeed it is impossible to choose a profile of the two quantities β_0 and U such that the three coefficients in (5.7) are constant. Even if the equation were made linear in ζ the interesting cases would have to be simulated by introducing discontinuities, and by considering two or more layers. The conditions that the pressure and vertical displacement be continuous across the interfaces would still be non-linear.

6. Bernoulli's equations for steady adiabatic motion

The equations of motion can be written in the form

$$\text{grad}\left(gz + \frac{q^2}{2}\right) - \mathbf{v}_\Lambda \,\text{curl}\,\mathbf{v} - \frac{1}{\tau}\,\text{grad}\,\varpi = 0 \quad . \quad . \quad . \quad (6.1)$$

If the velocity or the temperature is discontinuous across an interface, since (6.1) holds everywhere the velocity is involved non-linearly; grad gz is constant. Since ϖ is continuous, if grad ϖ is discontinuous at an interface it can only be in the component perpendicular to the interface. If the motion is not horizontal then there will be a gradient of velocity along streamlines and if the velocity is discontinuous across an interface then both grad $\frac{q^2}{2}$ and curl $\mathbf{v}$ are infinite there.

Since τ is constant along a streamline when the motion is adiabatic the component of Eq. (6.1) in the direction of motion gives

$$\frac{D}{Dt}\left(gz + \frac{q^2}{2} - \frac{\varpi}{\tau}\right) = \mathbf{v} \,.\, \text{grad}\left(gz + \frac{q^2}{2} - \frac{\varpi}{\tau}\right) = 0 \quad . \quad . \quad (6.2)$$

Thus $\left(gz + \frac{q^2}{2} - \frac{\varpi}{\tau}\right)$ is constant along a streamline, and in particular along the streamlines an infinitesimal distance on either side of an interface. If τ is discontinuous at the interface, (discontinuity of temperature, or density) since ϖ and z are continuous q^2 is discontinuous by an amount which varies as $\left(gz - \frac{\varpi}{\tau}\right)$ varies along the streamline. If τ is continuous and the velocity is discontinuous at one point it is discontinuous all along the interface by an amount which varies along the streamline.

7. The perturbation equations

By considering wave motion in a stream of incompressible fluid composed of layers in each of which density was uniform and the flow irrotational, and by letting the number of layers tend to infinity it was shown using linearized equations (Scorer 1951, Eq. (11)) that the equation for the vertical displacement assumed the form

$$\nabla^2 \zeta + \left(\frac{2U'}{U} + \frac{\rho_0'}{\rho_0}\right)\frac{\partial \zeta}{\partial z} - \frac{g\rho_0'}{U^2\rho_0}\zeta = 0 \quad . \quad . \quad . \quad (7.1)$$

which is Eq. (5.7) with the terms in $(\nabla\zeta)^2$ and g/c^2 (i.e. in ϵ) omitted. In the meteorological case, as we saw in Section 5, we may ignore the terms in ρ_0'/ρ_0 (i.e. in β_0) except where multiplied by g because the terms are small compared with terms in ϵ which themselves only have a small effect.

In applying Eq. (7.1) it is assumed that the coefficients are functions of z only, whereas in fact they are constant along streamlines and therefore vary with x when the motion is not horizontal. If the vertical displacements are large, the assumption that they are independent of x is only valid when the velocity and static-stability profiles are related in a particular way, viz. $U'/U = \text{const.}$, $g\beta_0/U^2 = \text{const.}$ Though realistic as a particular case, this does not provide a basis for general discussion. The form of Eq. (5.8) corresponding to Eq. (7.1) is

$$\nabla^2 \zeta + \frac{2U'}{U}\frac{\partial \zeta}{\partial z} + \frac{g\beta_0}{U^2}\zeta = 0 \quad . \quad . \quad . \quad . \quad (7.2)$$

The perturbation equation is therefore virtually correct if the coefficients are regarded a constant along a streamline, and in this light its accuracy can be assessed when it assume them to be constant along horizontal lines.

Since z_0 is constant along a streamline, and $z_0 = z - \zeta$

$$-u\frac{\partial\zeta}{\partial x} + w\left(1 - \frac{\partial\zeta}{\partial z}\right) = 0 \quad . \quad . \quad . \quad . \quad (7.3$$

The linearized version of this is obtained by writing $u = U$, and ignoring $w\partial\zeta/\partial z$ to giv

$$w = U\frac{\partial\zeta}{\partial x} \quad . \quad . \quad . \quad . \quad . \quad . \quad (7.4$$

Differentiating Eq. (7.2) with respect to x and substituting for ζ in terms of w we obtai

$$\nabla^2 w + \left(\frac{g\beta_0}{U^2} - \frac{U''}{U}\right) w = 0 \quad . \quad . \quad . \quad . \quad (7.5$$

on the assumption that the coefficients are independent of x. Writing $\partial/\partial x = ik$, fo small perturbations, and

$$l^2 = g\beta_0/U^2 - U''/U \quad . \quad . \quad . \quad . \quad . \quad (7.6$$

this becomes

$$\partial^2 w/\partial z^2 = (k^2 - l^2)\, w \quad . \quad . \quad . \quad . \quad . \quad (7.7$$

and in this form many applications of the theory have been made to practical problem either by assuming that l^2 is constant in each of a finite number of layers or that it varie in such a way as to give w as a simple function of z. The assumption that l^2 is constan is not very restrictive as to the kind of velocity and stability profiles which can be use for calculations in illustrative examples. Realistic profiles can easily be obtained (Score 1949-54) by taking two or three layers : although l^2 is discontinuous at the interface neither velocity nor temperature need be.

8. Oblique two-dimensional motion

If a velocity, which may vary with z_0, is added in the direction perpendicular to th two-dimensional motion already discussed the shape of the streamlines is unaffecte The effect is simply to add a velocity component V, which is constant along a streamlin

Since the motion is two-dimensional $\partial/\partial y = 0$ and so

$$\frac{Dv}{Dt} = 0, \quad \text{and} \quad v = V(z_0) \quad . \quad . \quad . \quad . \quad (8.1$$

The other equations of motion do not contain V if the earth's rotation is ignored and s are identical with those obtained when $V = 0$.

9. Motion in three dimensions

The motion in three dimensions cannot generally be described by a statement that certain function is constant along a streamline, since the velocity depends on the natur of the motion in adjacent streamlines and this motion is not related to that in the streamlin under consideration in the same way at all points along it, as it is in two-dimensiona motion.

But if terms of the order of ϵ are ignored, when U and β_0 are independent of z_0 Eq. (5.7) for ζ is

$$\nabla^2 \zeta + \frac{g\beta_0}{U^2} \zeta = 0 \quad . \quad . \quad . \quad . \quad (9.1)$$

We might therefore wish to superpose the flow over a set of ridges inclined at different angles to the basic stream but intersecting in one point, in the neighbourhood of which their heights are additive, to simulate an isolated hill. But since each component produces variations of velocity along the length of those to which it is not parallel only perturbation theory, which ignores such variations, can be used, even though Eq. (9.1) shows that solutions in two dimensions are superposable in that special case.

If we wish to investigate the circumstances under which three-dimensional motion can produce large shear which does not exist in the basic airstream we calculate the variation of curl **v** along a streamline as follows :

The equation of motion is

$$\tau \frac{\partial \mathbf{v}}{\partial t} + \tau (\mathbf{v} \,.\, \text{grad})\, \mathbf{v} = - \text{grad}\, \varpi + \tau \mathbf{g} \quad . \quad . \quad . \quad (9.2)$$

of which the curl is

$$\text{curl} \left(\tau \frac{\partial \mathbf{v}}{\partial t} + \tau \,\text{grad}\, \frac{q^2}{2} - \tau \mathbf{v}_\Lambda \,\text{curl}\, \mathbf{v} \right) = \text{grad}\, \tau_\Lambda \mathbf{g} \quad . \quad . \quad (9.3)$$

and writing $\mathbf{g} = -\,\text{grad}\, gz$ this simplifies to

$$\frac{1}{\tau} \text{grad}\, \tau_\Lambda \left\{ \frac{\partial \mathbf{v}}{\partial t} + \text{grad} \left(\frac{q^2}{2} + gz \right) - \mathbf{v}_\Lambda \,\text{curl}\, \mathbf{v} \right\} + \text{curl} \frac{\partial \mathbf{v}}{\partial t} - \text{curl}\, (\mathbf{v}_\Lambda \,\text{curl}\, \mathbf{v}) = 0 \quad . \quad (9.4)$$

In two-dimensional motion, including the oblique flow case, $(\mathbf{v}_\Lambda \,\text{curl}\, \mathbf{v})$ is perpendicular to the isentropic surfaces so that grad τ_Λ $(\mathbf{v}_\Lambda \,\text{curl}\, \mathbf{v}) \equiv 0$. In three-dimensional motion the contribution from this term is small because the product of a velocity and a velocity gradient is small compared with g except when the shear is very large, in which case also the term is small, because in the case we are interested in both **v** and curl **v** are almost perpendicular to grad τ, i.e., the shear is between isentropic layers and not within them.

In the cases under consideration, that is when there is large shear, grad $\frac{q^2}{2}$ is a vector very nearly parallel to grad τ; furthermore, grad $\frac{q^2}{2} \ll$ grad gz since the accelerations are small compared with those in free fall under gravity so that this term may be neglected (this is equivalent to ignoring the terms in $(\nabla\psi)^2$ in Eq. (5.5)). If we also assume that $\partial \mathbf{v}/\partial t \ll \mathbf{g}$ then Eq. (9.4) is

$$\text{grad} \log \tau_\Lambda \,\text{grad}\, gz + \frac{\partial}{\partial t} \,\text{curl}\, \mathbf{v} - \text{curl}\, (\mathbf{v}_\Lambda \,\text{curl}\, \mathbf{v}) = 0 \quad . \quad . \quad (9.5)$$

which can be written

$$\frac{D}{Dt} \,\text{curl}\, \mathbf{v} = \left(\frac{\partial}{\partial t} + \mathbf{v} \,.\, \text{grad} \right) \text{curl}\, \mathbf{v}$$

$$= - \,\text{grad} \log \tau_\Lambda \,\text{grad}\, gz + (\text{curl}\, \mathbf{v} \,.\, \text{grad})\, \mathbf{v} \quad . \quad . \quad (9.6)$$

This tells us that the change of curl **v** following the motion is produced mainly by two causes :

(1) $-$ grad log τ_Λ grad gz indicates a change of rotation of the elements about t contours of the isentropic surfaces so as to make them horizontal if the air is statica stable, or to increase their tilt if it is unstable.

(2) (curl **v**. grad) **v** shows that the vorticity increases if there is a velocity gradie in the direction of curl **v**, i.e., if there is a stretching of the vortex lines.

The greatest shear is likely to be produced where the two effects are additive. The are two possibilities : if the wind increases with height lateral stretching increases th forward shear and so decreases the surface wind, while if the wind decreases upwar lateral stretching will increase the surface wind.

On the windward side of an isolated hill the wind generally increases upwards an the vortex lines are stretched as the air passes partly round the sides of the hill. Th surface wind is thus reduced by this mechanism on the windward-facing slope.

On the lee side of a hill when the air is stably stratified the effect of gravity is to produ vorticity in the sense of wind decreasing with height. Thus strong surface winds und less strong upper winds are often found on lee slopes and under lee-wave troughs. the configuration of the ground is such as to produce sideways stretching in these plac the already strong surface wind will be further increased. On the other hand, if the a converges from either side in such a region, strong surface winds may be reduced : fo instance at the bottom of a converging valley on a lee slope the surface wind will probab not be increased by the convergence.

If vorticity about an axis in the direction of flow is produced on the side of a hill b the tilting of the isentropic surfaces about this axis and the air is then accelerated dow a lee slope the vortex lines will be stretched and this may produce a strong lateral compone of surface wind.

These principles can be applied in explaining the strong winds which are often e disastrous to trees on the lee side of a mountain. The effects will only be produced t a marked degree if the air is statically stable.

When the two effects are opposed the effect is to reduce any wave effects in com parison with the two-dimensional case.

10. Instability at an inversion

Finally a cursory investigation of the possibility of instability at an inversion ma be made. Eq. (5.8) is equivalent to

$$\frac{D}{Dt}\left(\text{curl } \mathbf{v} + \frac{g\beta_0}{U}\zeta\right) = 0 \qquad . \qquad . \qquad . \qquad (10.1$$

from which it is seen that along a stream-line on which β_0 is very large (large gradient of θ, large shear is generated in antiphase with ζ.

If the inversion is very pronounced grad log τ will be large and so *in a given wa form* greater vorticity will be produced than with a smaller value of grad log τ. Th vorticity created is rotation about the contours of the isentropic surfaces and is increase as described above on the windward or leeward side of a hill by lateral stretching of the ai and on the side of a hill by longitudinal stretching. The amount of vorticity introduce by gravity is proportional to the static stability, because $|\text{grad log } \tau| = \beta$. Thus if u denotes the shear we have $u' \propto \beta$. But the Richardson number is proportional to β/u'^2 and so if we take the value of this number as a criterion whether the shearing flow wi become unstable, the likelihood of this happening is increased if β is increased whatev the critical value may be. The greater the static stability of a layer the greater is th tendency to produce dynamic instability in mountain waves.

The argument so far assumes that the wave form is given. But it has been shown (Scorer 1954) that in two-dimensional flow the maximum of the wave amplitude is usually more pronounced if there is an inversion than if there is not, this maximum being at the inversion. Thus, although no general theorem can be stated it can be said that inversions do not usually serve to decrease the slope of the streamlines in a wave but more often increase it in comparison with other cases. The conclusion that the greater the stability the greater is the likelihood of clear-air turbulence in mountain waves is therefore reinforced.

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Gust variation with height up to 150 m

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SUMMARY

Measurements, on occasions of strong wind, of horizontal wind gusts at heights of 12, 64 and 153 r in open country show the increase of gust speeds with height to be markedly less than that of mean win speed. Lack of synchronism between gusts at the various heights makes it necessary, for wind-pressu applications, to examine the maximum values of wind speed averaged over the three heights. At the time of occurrence of these maxima, the wind speed is approximately proportional to the height raised to th power 0·085. The corresponding index for the mean speed is 0·16.

Some results on the structure of turbulence are applied to enable predictions of extreme vertical gus velocities to be made.

1. INTRODUCTION

In the design of radio masts and other high structures an important consideration i the maximum wind pressure which must be taken into account. To be able to comput this it is necessary to know, not only the maximum surface wind to allow for, but als the variation of wind speed with height. Climatological records for the area in question may provide means of estimating the former but information on the variation with heigh is still meagre. Some of the formulae proposed at various times have been based on th normal variation of *mean* wind speed with height but this is unsatisfactory. It is reasonabl to suppose that strong gusts during storms are the result of the descent of relatively large bodies of air carrying with them a high wind speed from aloft and it is to be expecte that the variation of gust speed with height should be less than that of mean speed a the latter reflects the average frictional retardation experienced by air masses (eddies) o very diverse size : small descending air masses rather quickly lose their store of momentum because of their relatively large ratio of surface to volume, i.e. friction to inertia.

Sherlock (1952) has investigated storm-gust structure at heights up to 250 ft bu as radio masts are often much higher it seemed useful to obtain data for heights up t 500 ft and these are presented in the present paper.

2. SITE AND METHOD

Anemometers were mounted at heights of 12·2 m (40 ft), 64 m (210 ft) and 153 m (503 ft) on a radio mast near Sale, Victoria (38°06′S, 147°04′E). The mast is on a slight rise amid gently rolling grazing land and for winds between N and W, such as prevailed on nearly all occasions when observations were made, the fetch of the wind is over open grassland with but few trees. The sea is distant some 20 mi to the south. The anemometers were mounted on outriggers from the mast and had a free exposure for winds from S through W and N to NE. No records were utilized for other wind directions.

The wind records were obtained using cup anemometers with Sheppard (1940) type rotors actuating electric contacts at a rate of approximately one contact per 5 m run of wind. The electric impulses were recorded by a 4-pen chronograph with a paper speed of 5 cm/sec. The fourth pen was connected to a contacting clock giving impulses each half second. This system has good resolving power so that mean wind speeds were evaluated for each 2-sec interval of the records. The analysis is laborious but the good

accuracy readily achieved with this system is a major consideration. A 2-sec analysis interval was chosen in consultation with radio-mast designers who considered very transient gusts to be unimportant.

Cup anemometers have usually been considered unsuitable for studies of wind gustiness but it can be shown as a result of Schrenk's (1937) treatment of cup-anemometer response that in a study of maximum gust speeds in strong winds they are capable of high accuracy. In terms of Schrenk's dimensionless parameter K (which involves air density, certain aerodynamic characteristics of the cup rotor and its moment of inertia, as well as mean wind speed (V_0) and the period (T) of the wind oscillations), Sanuki (1952) has shown that the cup anemometer gives full response to the wind maxima in a sinusoidally fluctuating wind when $K > 2$. This was independently found, before the present work was commenced, by numerical integration of Schrenk's Eq. (2a) which gave the results shown in Table 1. A is the amplitude of the wind oscillation as a fraction of the mean wind speed, V_i (max)/V (max) is the indicated wind maximum as a fraction of the true maximum and ϕ the phase lag of the indicated maximum.

TABLE 1. CUP-ANEMOMETER RESPONSE TO WIND MAXIMA CALCULATED FOR A SINUSOIDAL FLUCTUATION

K	A	$V_i\,(max)/V\,(max)$	$\phi°$
2·0	0·5	0·998	9
1·5	0·5	$0·991_5$	13
1·5*	0·5	$0·988_5$	13
1·0	0·5	0·980	20·5
1·0	0·33	0·982	22
1·0	0·25	0·984	23·5

* Solution including a bearing frictional torque of 1,000 dyne cm in a mean wind speed of 10 m/sec. The anemometers used had a friction considerably less than this.

It is apparent from Table 1 that the underestimation of the wind maxima is small for $K \geqslant 1·5$. For the Sheppard rotors used in the present work $K = 0·031\,V_0\,T$, with V_0 in m/sec and T in seconds (Deacon 1951). Thus for mean wind speeds of 10 m/sec or more we have $K > 1·5$ for all values of $T > 5$ sec and, as analysis of the records in terms of 2-sec mean speeds only takes account of fluctuations of periods greater than 5 sec, it follows that the gust speeds dealt with below are almost completely unaffected by instrumental inertia or friction.

3. NOTATION

$u\,(z)$ = wind speed at height z ft

$\bar{u}\,(z)$ = mean wind speed at height z ft during a given observation period

$\bar{u}_{10}$ = mean wind speed at 10-m height

$u_e\,(z)$ = maximum $u\,(z)$ during a given period

$u_e'\,(z) = u_e\,(z) - \bar{u}\,(z)$

$u\,(m) = \frac{1}{3}\{u\,(40) + u\,(210) + u\,(503)\}$, where the wind speeds within the brackets are simultaneous values

$u_e\,(m)$ = maximum value of $u\,(m)$ in a given period

α = index of z in power-law variation of mean wind speed with height

$\alpha\,(e)$ = index of z in power-law variation of u_e with height

4. Results

Altogether 31 successful observations over periods of 5 to 15 min were obtained on occasions of strong wind ($>$ 20 mi/hr at 40 ft) and analysed for mean wind speeds over each 2-sec interval in each case. The results derived from these are set out in Table 2 while portions of the records for two occasions are given in Fig. 1; the occasions were those on which gusts of just over 70 mi/hr were recorded at 503 ft.

TABLE 2. Results of observations of wind speeds at heights of 40, 210 and 503 ft on a radio mast at Sale, Victoria

Obsn. No.	Date 1952	Time	Durn. (min)	$\bar{u}$ (40) (mi/hr)	$\bar{u}(210)/\bar{u}(40)$	$\bar{u}(503)/\bar{u}(40)$	Wind speed maxima (mi/hr) $u_e(40)$	$u_e(210)$	$u_e(503)$	$u_e'/\bar{u}$ at 40 ft	210 ft	503 ft	$u_e(m)$ (mi/hr)
	Sept.												
2	2	1917	8·9	27·6	1·34	1·70	36·0	48·0	54·5	0·30	0·30	0·16	42·8
3	14	1406	6·6	27·3	1·22	1·39	37·0	42·0	45·0	0·36	0·26	0·18	40·2
4	14	1634	8·6	30·6	1·25	1·54	43·5	51·5	54·5	0·42	0·35	0·16	48·7
5	16	1615	8·5	30·6	1·30	1·58	44·5	52·5	60·0	0·45	0·32	0·25	47·7
6	16	1715	5·8	24·7	1·27	1·55	34·0	36·5	45·0	0·38	0·17	0·15	37·2
7	16	1821	4·9	40·8	1·32	1·51	64·0	67·0	71·0	0·57	0·24	0·17	64·3
8	16	1828	8·4	31·3	1·33	1·78	48·0	56·0	66·0	0·53	0·35	0·19	52·1
9	16	1920	7·8	28·2	1·33	1·55	43·0	51·0	55·5	0·53	0·36	0·27	44·7
10	17	1327	9·0	34·7	1·18	1·39	44·0	53·0	59·0	0·27	0·29	0·22	48·8
11	17	1427	4·4	29·0	1·27	1·58	38·5	46·0	51·5	0·33	0·24	0·12	41·7
12	17	1925	9·3	21·1	1·37	1·77	31·0	42·5	48·5	0·47	0·48	0·30	35·6
13	17	2015	6·8	27·0	1·31	1·60	40·5	45·5	56·5	0·50	0·29	0·31	46·0
14	17	2155	13·4	21·6	1·30	1·68	34·5	37·5	51·0	0·60	0·34	0·40	38·5
15	17	2330	7·6	28·5	1·29	1·70	39·0	47·0	53·0	0·37	0·38	0·19	43·8
16	19	0940	10·3	26·6	1·19	1·37	34·5	41·5	45·0	0·30	0·31	0·24	36·9
17	19	1130	7·5	25·1	1·20	1·36	39·0	45·0	45·0	0·55	0·47	0·31	39·6
18	20	1000	8·5	27·0	1·26	1·53	38·5	44·5	49·0	0·43	0·31	0·18	42·1
19	20	1020	7·1	28·8	1·14	1·38	44·0	42·0	49·0	0·53	0·28	0·23	42·8
20	20	1140	6·3	27·5	1·23	1·33	38·5	44·5	47·5	0·40	0·32	0·30	42·3
21	20	1335	3·7	32·8	1·18	1·39	45·5	47·0	58·0	0·39	0·21	0·27	47·9
22	20	1443	9·1	27·1	1·14	1·26	38·0	39·0	45·0	0·40	0·26	0·31	38·2
	Oct.												
23	15	2148	13·2	23·2	1·34	1·58	38·5	45·5	51·5	0·66	0·46	0·40	42·6
24	15	2248	15·2	27·6	1·35	1·76	43·0	51·0	61·0	0·56	0·37	0·26	48·8
25	16	1100	4·5	24·5	1·19	1·38	34·0	37·0	44·5	0·39	0·27	0·31	38·4
26	17	0130	7·2	22·3	1·39	1·92	32·0	39·5	50·0	0·44	0·28	0·17	37·9
27	17	0407	5·1	26·0	1·40	1·79	36·5	45·0	55·5	0·40	0·24	0·19	41·6
28	17	1420	8·8	21·8	1·20	1·40	31·0	32·0	38·5	0·42	0·22	0·27	30·8
34	28	1450	8·0	24·6	1·25	1·40	35·5	38·5	43·0	0·44	0·26	0·25	39·0
	Nov.												
36	2	1420	8·3	25·1	1·23	1·40	34·5	39·5	41·0	0·38	0·28	0·17	37·3
37	10	1720	12·2	25·7	1·32	1·54	45·5	51·0	64·5	0·77	0·50	0·63	52·5
38	19	1330	10·2	35·2	1·31	1·59	55·0	58·5	70·0	0·56	0·27	0·26	59·0
Means			8·2	27·6	1·272	1·540	Geometric means			0·443	0·303	0·237	—
Standard deviations			—	—	0·070	0·162	St. devn. of $\log_{10}(u_e'/\bar{u})$			0·102	0·106	0·148	—

It will be seen from Table 2 that with only one exception (Run 19), the values of $u_e(z)$ progressively increase with height. The average extent of this increase is shown by the mean ratios of gust maxima given in Table 3.

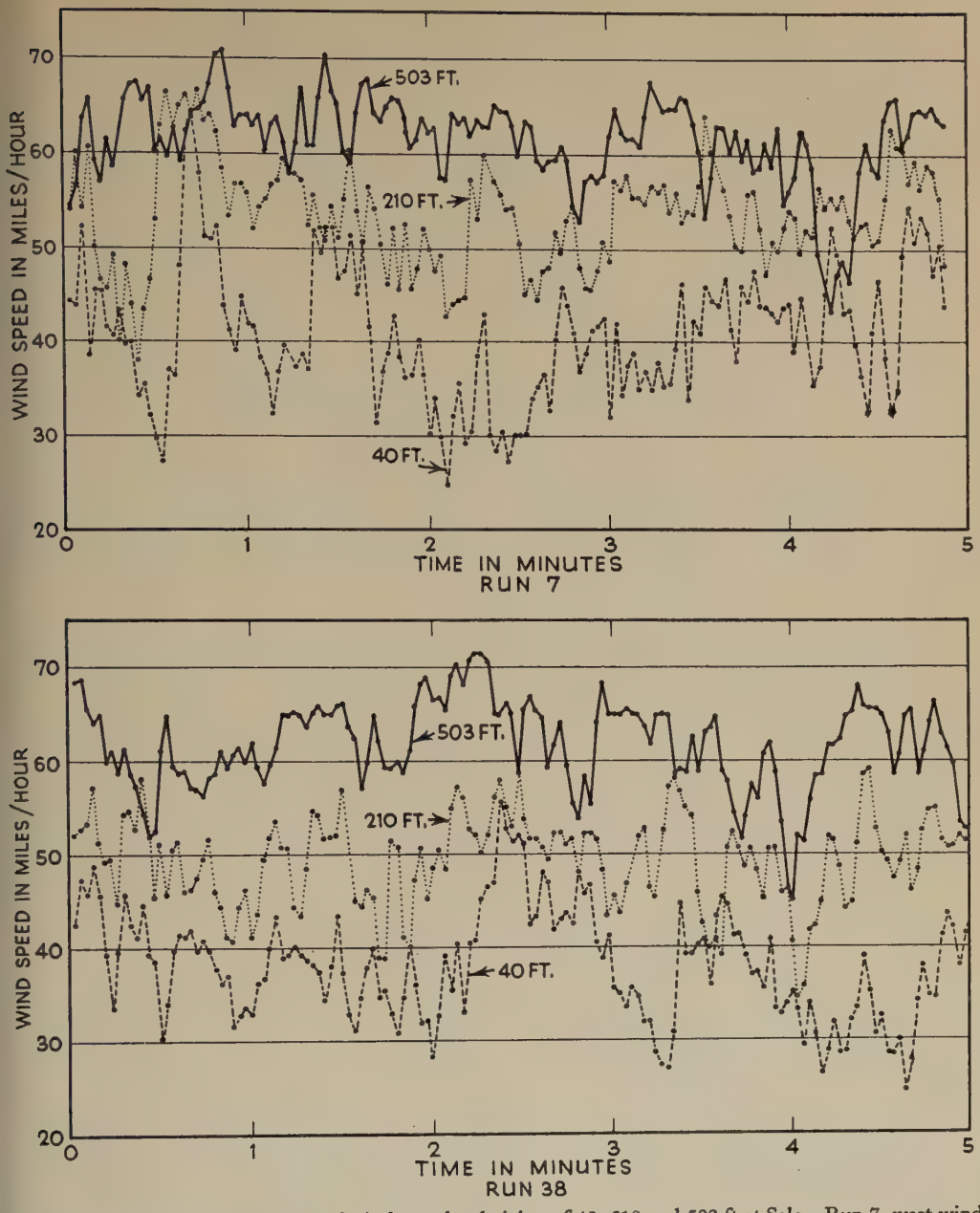


Figure 1. Specimen observations of wind speed at heights of 40, 210 and 503 ft at Sale. Run 7, west wind; Run 38, north wind.

TABLE 3. AVERAGE VALUES OF GUST-MAXIMA RATIOS

u_e (40) (mi/hr)	30-34	35-39	40-44	45-65	35-65
Mean u_e (210)/u_e (40)	1·156	1·164	1·150	1·092	1·145
Standard deviation	0·12	0·075	0·085	0·055	0·078
Mean u_e (503)/u_e (40)	1·394	1·292	1·313	1·294	1·298
Standard deviation	0·13	0·12	0·105	0·12	0·111
Number of observations	7	12	7	5	24

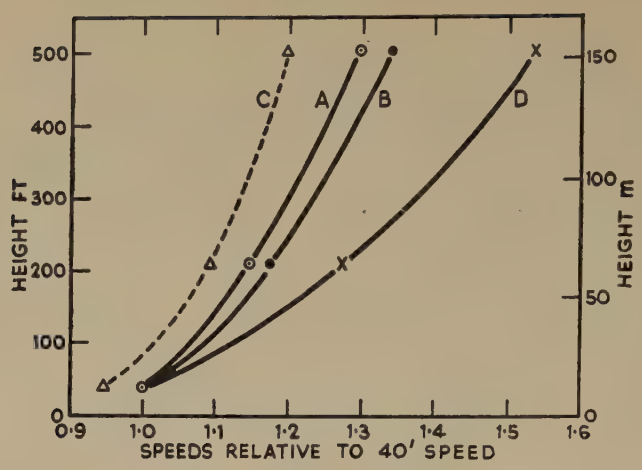


Figure 2. Height variation of gust speeds and of mean wind speed. Curve A – Based on maximum gusts in average period of 8 min irrespective of times of occurrence at the various heights. Curve B – As 'A' but for 2-min periods. Curve C – Average height variation relative to 40-ft maximum gust, of the speeds occurring at instants when $\frac{1}{3}(u_{40} + u_{210} + u_{503})$ has maximum value in each run. Curve D – Variation of mean wind speed with height; average of all runs.

Table 3 shows that the mean gust ratios are effectively constant for 40 ft gust speeds of 35 mi/hr or more. Below 35 mi/hr the 503 : 40-ft ratio is larger by an amount just significantly greater than would be expected from the observational scatter. The means for the range 35-65 mi/hr given in the last column of Table 3 may most suitably be adopted as applicable to storms and they are plotted in Fig. 2, curve A.

For these same 24 occasions another set of gust ratios were obtained with the highest gust speeds in each two minutes of record taken into account instead of only the strongest gust in each run. The mean values so obtained are :

$$210/40\text{-ft ratio} = 1{\cdot}174$$

$$503/40\text{-ft ratio} = 1{\cdot}341$$

These ratios (shown in Fig. 2, curve B) are larger than those above and this suggests that the rarer more outstanding gusts give generally a smaller variation with height than the rest, a result intelligible in terms of the effect of gust scale on the preservation of momentum by a gust in its descent.

That the strong gusts do, in general, commence earlier at the greater heights than at 40 ft is shown by a detailed examination of the records.* For two runs the correlation coefficients between wind speed at 210 ft and that at 40 ft, t seconds later, are shown in Fig. 3 for a range of t from $+15$ to -15 sec. It is apparent that, under these conditions of mean 40-ft wind speed 35 to 40 mi/hr, the correlation reaches a maximum of about 0·5 when t is of the order of 5 sec. The values relating to the 503-ft wind, also given in Fig. 3, show a marked decrease in the magnitude of the correlation with increase in the height interval; for Run 7 the 503 : 40-ft coefficients are too small to be significant. This indicates that the majority of eddies are too small in scale or lifetime for them to be able to affect both levels when the vertical separation is as great as 460 ft. Usually the more marked gusts may be distinguished in the records at all three heights. These however are too infrequent to contribute much to the overall correlation between speeds at 503 and at 40 ft.

* Blue-print copies of these in a form similar to Fig. 1 can be supplied on request.

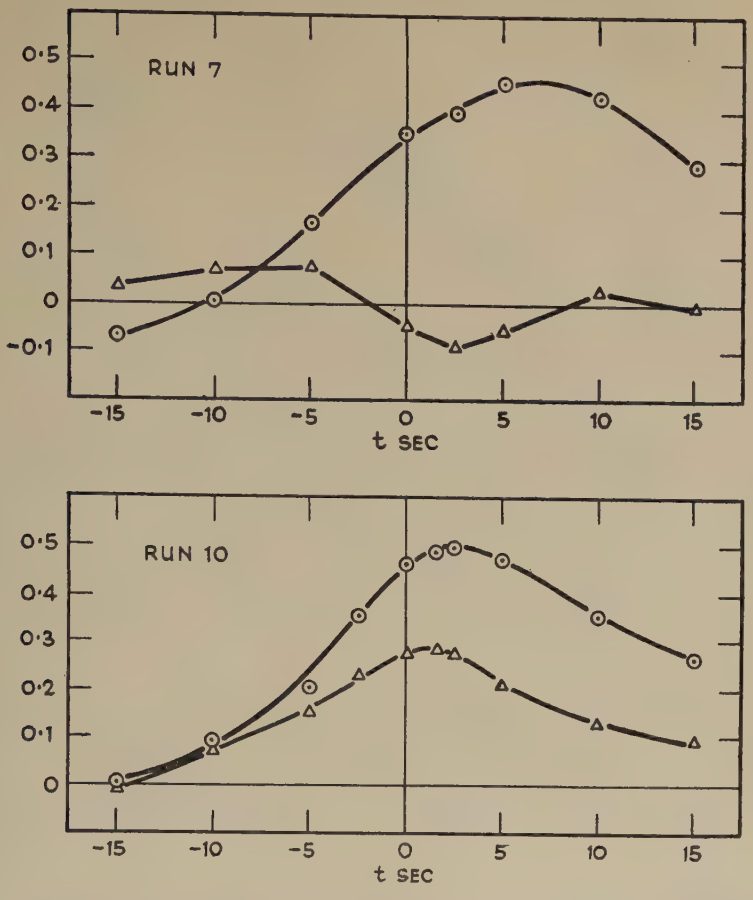


Figure 3. Correlation coefficients between wind speed at 503 ft or 210 ft and that at 40 ft, t sec later.
⊙ 210 and 40 ft values, △ 503 and 40 ft values.
Run 7, $\bar{u}\,(40)$ 40·8 mi/hr, RMS values (mi/hr) of u' : 7·1 (40 ft), 5·6 (210 ft), 4·55 (503 ft).
Run 10, $\bar{u}\,(40)$ 34·7 mi/hr, RMS values (mi/hr) of u' : 5·6 (40 ft), 4·75 (210 ft), 4·05 (503 ft).

The earlier onset of strong gusts at the heights greater than 40 ft is probably a combined effect of descent of air masses and of sloping gust fronts of the type discussed by Sherlock (1952). As vertical velocity components were not measured at Sale it is impossible to disentangle these effects. But some records of both horizontal and vertical wind-components at a height of 28 m at Edithvale, Victoria, show that about 75 per cent of the stronger gusts are accompanied by downward motion so that the latter is, as would be expected, an important element in gust mechanics.

The lack of synchronism in the gusts at the several heights makes it necessary, for practical purposes, to examine the maximum values of wind speed averaged over the height of the mast in relation to the gust maxima at 40 ft, $u_e\,(40)$. The maximum value of $u\,(m) = \frac{1}{3}\{u\,(40) + u\,(210) + u\,(503)\}$ for each run is accordingly listed as $u_e\,(m)$ in the last column of Table 2. For wind-pressure studies the RMS value of the three wind speeds would be more significant than the mean but it was found by trial of a few cases that the difference is small and the considerable additional labour in finding the RMS values not warranted. The mean ratio for all runs is

$$u_e\,(m)/u_e\,(40) = 1{\cdot}089$$

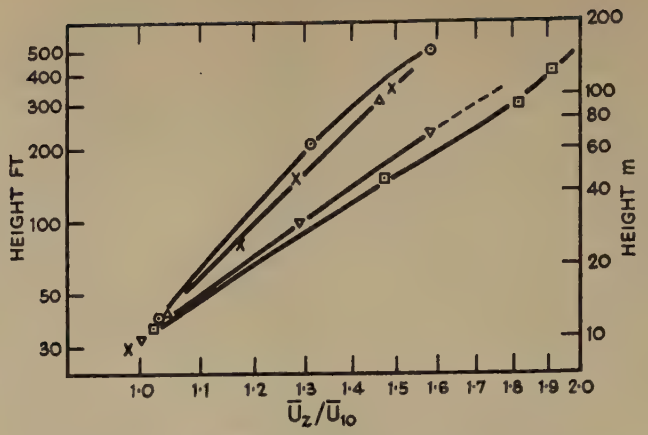


Figure 4. Increase of mean wind speed with height.
⊙ Sale, × Cardington, △ Leafield, ▽ Quickborn, ⊡ Upton, Long Island, U.S.A.

with a standard deviation of the individual values of 0·055. This ratio is appreciably smaller than the value 1·164 which would be obtained from Table 2 on the assumption that the gust maxima were simultaneous at all heights.

The variation of wind speed with height at the times when $u\,(m)$ is a maximum is shown in Table 4 and Fig. 2 (curve C).

TABLE 4. RATIOS OF WIND SPEED AT 40, 210 AND 503 FT, WHEN THE SUM OF THE THREE IS A MAXIMUM, TO THE 40-FT GUST MAXIMUM : MEANS FROM 31 RECORDS

Height (ft)	40	210	503
Ratio	0·947	1·092	1·197

The variation of mean wind speed with height is of interest for comparison with the gust maxima and the mean values of $\bar{u}\,(210)/u\,(40)$ and $\bar{u}\,(503)/\bar{u}\,(40)$ given in Table 2 are shown in Fig. 2, curve D. There is no significant variation of the mean wind ratios with wind speed over the range experienced.

5. DISCUSSION OF RESULTS

(a) *Variation of mean wind speed with height*

The variation of mean wind speed with height in these Sale observations is first compared with results for other sites as this gives some information on the aerodynamic character of the terrain. Furthermore, in the absence of anomalous features in the mean wind profiles, more confidence may be placed in the vertical variation of the gust velocities as being reasonably free from purely local peculiarities. The comparison is shown in Fig. 4, the winds at the various heights being expressed in ratio to that at 10 m height. Where not directly measured the 10-m wind speed ($\bar{u}_{10}$) has been obtained by graphical interpolation. The results in all cases are either for strong winds such that thermal stratification and its effects are negligible or else for an effectively adiabatic lapse rate of temperature in which case the wind ratios should approximate closely to those observed in strong winds. Logarithmic scales are used for both height and wind-ratio axes as it has usually been found that a simple power-law expression gives a reasonably good representation of wind profiles in the range of heights here considered.

The observations at Cardington, Bedfordshire (Frost 1947) were made using an anemometer suspended from a barrage balloon. The surrounding country is relatively flat agricultural land with but few trees. The variation of mean wind speed with height given by 57 observations for approximately adiabatic conditions is as follows :

Height (m)	1·2	9·2	24·5	46	106
$\bar{u}_z/\bar{u}_{10}$	0·69	0·98	1·17	1·28	1·49

At Leafield, Oxfordshire, observations were made on a radio mast standing on a broad ridge of the Cotswolds (Heywood 1931) and the surroundings were mainly open fields divided by low stone walls and hedges. In Heywood's Table 7 there are 38 observations with approximately adiabatic lapse rate and wind speed at 95 m greater than 8·5 m/sec and these give a mean ratio $\bar{u}_{95}/\bar{u}_{12\cdot7} = 1{\cdot}40$. Fig. 4 shows this variation with height to be somewhat greater than at Cardington, which is what would be expected from the descriptions of the sites.

Some results for Quickborn near Hamburg (Franckenberger and Rudloff 1949) are for occasions of wind speed at 70 m $\geqslant$ 15 m/sec in January. The mean of 21 observations gives :

Height	10	30	70
$\bar{u}_z/\bar{u}_{10}$	1·00	1·285	1·508
Standard deviation of observations	—	0·067	0·074

The variation with height is greater than for the sites previously mentioned and this is consistent with the nature of the site, which is relatively level meadow land but with numerous hedges and trees around the small fields making the terrain very rough aerodynamically. That Quickborn is a considerably rougher site than Nauen near Berlin is shown by the following comparison of annual averages given by Franckenberger and Rudloff (1949) using Hellmann's material for Nauen.

	10 m	30 m	70 m
Annual average (m/sec) at Quickborn	3·6	5·2	7·0
Annual average (m/sec) at Nauen	4·3	5·2	6·3

A few observations are available for another rough site at Upton, Long Island, N.Y. (Panofsky 1953) where the terrain, generally flat or slightly rolling, is largely covered with patches of scrub oak and scrub pine about 10 m in height. The mean of 8-hr recording under conditions of 91-m wind $>$ 9 m/sec gives :

Height (m)	11	45	91	125
$\bar{u}_z/\bar{u}_{10}$	1·02	1·47	1·81	1·91

Fig. 4 shows that, of the localities considered, Upton has the greatest variation of wind with height, and this again accords with the nature of the site and with the magnitude of the shearing stress (τ) measured at 91-m height during these observations. The geometric mean value of the coefficient k in $\tau = k\rho\bar{u}_{10}^2$ is 0·015, a high value by comparison with that of about 0·005 for open grassland.

Over the height range 10 to 100 m there is no definite or consistent departure from linearity to be discerned in the graphs of Fig. 4. The simple power law

$$\bar{u}_z/\bar{u}_a = (z/z_a)^\alpha$$

where $\bar{u}_a$ is the wind speed at reference level z_a, is therefore a useful representation of the mean wind profile in this height range under conditions of strong wind or adiabatic lapse rate. This being so, the values of α and of $\bar{u}_{100}/\bar{u}_{10}$ are given in Table 5.

TABLE 5. VALUES OF INDEX α IN THE SIMPLE POWER-LAW EXPRESSION FOR THE VARIATION OF MEAN WIND SPEED WITH HEIGHT. CONDITIONS: STRONG WIND OR ADIABATIC LAPSE RATE; HEIGHT RANGE 10 TO 100 m

Site	$\bar{u}_{100}/\bar{u}_{10}$	α
Sale, Victoria	1·44	0·16
Cardington, Beds.	1·48	0·17
Leafield, Oxfordshire	1·48	0·17
Quickborn, Germany	1·71	0·23
Upton, Long Is., U.S.A.	1·83	0·26
Akron, Ohio, U.S.A.	—	0·22

The last entry in Table 5 is a mean value obtained from the results of Huss and Portman (1949) for a site in gently rolling country with many bushes and small trees. These authors evaluated α by least squares from 5-min mean velocities at six heights between 12 and 107 m. The 40 values for wind speeds greater than 9 m/sec at 50 m give a mean $\alpha = 0{\cdot}219$ with a standard deviation of the individual values of 0·042.

From this comparison of results it appears that the mean wind profile at Sale exhibits no unusual features and is typical of level and practically treeless grassland sites.

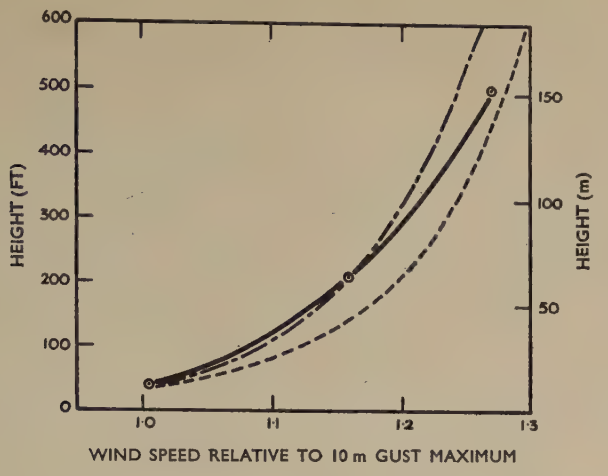
(*b*) *Variation of gust maxima with height*

The most thorough study of storm gust velocities with which the present results may be compared is that of Sherlock (1952). Special quick-response anemometers were mounted at 25-ft intervals up to 250 ft (76 m) on a mast in open country and numerous records obtained during winter storms over a period of seven years. Sherlock finds a value of α, the power-law index for the variation of mean wind speed with height, of about 0·14. This is reasonably close to the value of 0·16 from the Sale data and indicates the sites to be fairly comparable.

As in the present work, Sherlock also finds the increase of gust maxima with height to be less rapid than of mean speed. If we denote by $\alpha\,(e)$ the index of z in a power-law variation of gust maxima with height, then Sherlock finds for 10-sec gusts that $\alpha(e) = 0{\cdot}080$. He selects the 10-sec gust for most attention for the reason that shorter-duration gusts are unlikely to envelop the whole height of a structure 200 or more feet high. His result is therefore most nearly comparable with curve C of Fig. 2 which gives $\alpha\,(e) = 0{\cdot}088$ for the 40 to 210-ft interval and 0·104 for the higher interval 210-503 ft. Remembering that Sherlock's observations only extended up to 250 ft the agreement is close.

For design purposes curve C of Fig. 2, as it stands, is not directly suitable as some allowance needs to be made for the fact that it is a mean curve; on about 50 per cent of occasions more severe conditions are experienced. The mean ratio $u_e\,(m)/u_e\,(40)$ of 1·089 (see Section 5) was derived from individual values having a standard deviation about this mean of 0·055. If therefore at each height the abscissae of curve C are increased by 5 per cent, a curve is obtained representing a condition of severity only likely to be exceeded on about one-sixth of occasions.* As the original observations had an average duration of just over 8 min, it follows that, in about 1 hr of storm one gust may be expected to be as unfavourable as is represented by the adjusted curve C. This seems a reasonable provision to make as very strong winds are rarely of long duration. In Fig. 5 this adjusted

* A uniform 5 per cent addition is not entirely satisfactory but the nature of the problem is such that many more observations extending to very high wind speeds would be needed to warrant more elaborate treatment.



gure 5. Comparison of variations of wind speed with height recommended for use in the design of high structures.

——— Sale observations. – — – — Sherlock's (1952) recommendation. – – – – – American Standards Association curve.

ariation with height* is shown to agree quite closely with Sherlock's proposed curve p to 250 ft, his highest observational level, but above this the curves diverge and the ifference becomes appreciable at 500 ft.

Also shown in Fig. 5 is the wind-speed variation with height corresponding to the ressure variation (smoothed) quoted by Sherlock (1952) from 'American standard esign loads in buildings and other structures' (ASA, A58.1, 1945). For the lowest 300 ft ie ASA curve gives a too rapid variation of wind speed with height to be suitable for pen grassland exposures.

The Sale curve of Fig. 5 is represented reasonably well (within 1 per cent at 210 nd 503 ft) by the equation

$$u_z/u_e\,(10\text{ m}) = (z/10)^{0\cdot085}$$

here z is measured in metres.

There are very few other published observations of the variation of gust speeds ith height. Bouvier (1950) has presented a few observations for heights of 39 m and 50 m at Sainte-Assise, France and discussed them in relation to some formulae (without eferences) for the height variation of wind pressure. The formulae of H. de la Noë and V. D. Tuck are both close to that for a wind speed proportional to height raised to a ower of 1/7, so it is likely that they were based, as was that of Wing (1921), on observations *mean* wind speed.

Bouvier stresses the fact that, on an occasion of strong gale, the wind maxima were n average equal at the two heights but gives no evidence to show whether or not the usts were simultaneous at the two levels. He then assumes the basic datum available a designer to be the maximum wind speed at 250 m height and concludes that it is nsafe to allow for any reduction in wind pressure at the lower levels. As however it more commonly the 10 m data from which a designer must work, it is not justifiable take Bouvier's conclusion to imply that no increase of wind with height should be lowed for in this latter case.

* Using curve B of Fig. 2; this has been further adjusted by another 1 per cent to make it applicable to the standard reference ght of 10 m (33 ft).

Carruthers (1943) quotes a series of gust ratios for heights up to 100 m which we obtained by E. G. Bilham (unpublished) from observations of J. S. Dines at Pyrton Hil Bucks. This height variation does not differ significantly from that indicated in Fig. for the Sale data.

The observations so far discussed relate to open grassland with few trees. As th influence of the vegetative cover of the ground has been shown to be quite well mark for the mean-velocity profiles, it is relevant to consider what may be the effect on th gust profiles. Observational guidance on this point is lacking but it seems very probab that a greater vertical gradient of mean velocity is accompanied by a steepening of th gust profile, probably roughly in proportion. Considering broader aspects of the proble however it seems unnecessary to attempt to take account of this. If an estimate is availab from climatological records of the highest 10-m gust likely to be experienced, in a give term of years, over an open site in the area considered and an appropriate allowan made for increase with height, then the conditions over an adjacent area of scrub woodland may safely be assumed to be somewhat less severe. The greater variation wi height over the rougher surface will be a consequence of the reduced gust maxima 10 m; at heights of 1,000 ft or so the gusts will be but little affected by the change surface. The lengthening series of observations from anemometers at airfields mak the choice of an open grassland exposure the most appropriate to adopt as standard. Ci observatory records extending back over longer periods may also be utilized if adjust by comparison with airfield data in the same district for the overlapping period. some cases varying exposure in different directions may be a complicating factor.

(*c*) *Vertical gust velocities*

Measurements of vertical gust velocities were not made at Sale but other results o the structure of turbulence enable us to infer the probable magnitude of vertical gus from the present data. This is of interest in the design of certain radio towers carryin large horizontal areas.

The influence of the ground causes the vertical eddy velocities near the ground t be distinctly smaller than the horizontal but G. I. Taylor (1927) first showed there t be a marked tendency for these to become equal at greater heights (of the order of 100 ft Subsequent work has supported this, for example Panofsky (1953) and Deacon (1955 and L. F. G. Simmons (personal communication) said of observations made with speciall designed equipment mounted at a height of 515 ft (157 m) on a mast at Blackpool :

> From a detailed analysis of one of the records – exhibiting the variations present in a wind average speed 15·5 m/sec during a period of one hour – we came to the conclusion that the large horizontal gusts, though often differing in shape from vertical gusts and of longer duration, were rough of the same intensity.

Calling the extreme vertical and horizontal eddy velocities in a given period, w_e' and u_e respectively, then it may be concluded that there is a fair weight of evidence to show tha $w_e' \simeq u_e'$ at heights above about 50 m. Simmons found approximate equality in intensit and frequency of upward and downward gusts but that the stronger vertical gusts ar of short duration, from 0·3 to 3 sec, only about one third that of the horizontal gusts.

Values of $u_e'/\overline{u}$ are given in Table 2 together with geometric means from the whol series of observations. Geometric means are given as the distribution of $u_e'/\overline{u}$ is skew on a logarithmic scale there is no significant departure from normality.

On the assumption that $w_e' = u_e'$, the geometric mean value $u_e'/\bar{u} = 0{\cdot}24$ for 503 ft indicates that during a period of high wind, vertical velocities of 24 per cent of the mean wind at 503 ft may be expected to occur about once in 8 min on the average (8 min being the mean recording period in Table 2). From the standard deviation of the logarithms of the individual values in Table 2 it may be calculated that about once per hour vertical velocities of 34 per cent of the mean wind, or slightly more, may be expected to occur. Simmons found from a number of records of vertical velocity in high winds at 515-ft height at Blackpool a corresponding value of 35 per cent.

The 8-min extreme gust at 210 ft of 30 per cent of the mean wind at that height similarly leads to an hourly value of 40 per cent. Simmons found a higher value, about 50 per cent, at 150-ft height at Cardington but much of the difference may be an effect of the large building close to the mast at Cardington.

From the mean hourly value of $u_e/\bar{u}$ at 40 ft of 0·57 (from the mean 8-min value 0·44) it follows that a horizontal gust of 100 mi/hr at 40 ft (often taken as the design maximum) is likely to be experienced in an hour of gale with $\bar{u}\,(40) = 64$ mi/hr. From curve D of Fig. 3 we see that mean wind speeds of 82 and 99 mi/hr are to be expected at 210 and 503 ft respectively and so, from the above figures, the maximum vertical gusts are likely to be 33 mi/hr at 210 ft and 34 mi/hr at 503 ft.

Acknowledgments

Acknowledgments are made to the Postmaster General's Department for the provision of facilities at Sale for this work and to Mr. F. Chapple and others of the Sale Radio Station for taking the records. Mr. E. A. Cohen analysed many of the data and Miss Y. M. Mullins computed the solutions of the anemometer equation.

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The cloud-seeding trials in the central Punjab, July-September 19

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1. Introduction

Experiments to test the influence of the dispersal of salt particles from the groun on rainfall amount and distribution were carried out in the central Punjab betwee 16 July and 15 September 1954.

The dispersal of hygroscopic particles rather than freezing nuclei was determine by the high cloud-base temperature (between 20°C and 25°C) prevailing in the Punja during this season. Monsoon showers frequently fall from clouds whose tops do not reac the freezing level, and certainly not the level at which ice crystals are normally found, c can be made to appear, in clouds. The factor determining the choice of the site for thes trials was the natural concentration of salt particles in the atmosphere. During th previous two years samples of airborne dust had been taken twice weekly at five statio in west Pakistan, and analyzed for their content of hygroscopic particles. On the averag the surface air in the Punjab during the monsoon season contains 5-10 hygroscopic particle per cubic metre whose dry mass is greater than 10^{-9} g. There was therefore some hope tha we should be able to disperse particles of the same size range at a rate sufficient at leas to double the hygroscopic particle content of the atmosphere over an area large enoug for any subsequent increase or modification of rainfall to be detected by the existin raingauge network.

This would not have been possible, for instance, in lower Sind, where cloud-bas temperatures are higher, but where the natural salt-particle content is some 20-50 time higher than in the Punjab.

It was decided from the outset that the method adopted for the dispersal of sal particles should be capable of being employed cheaply and on a large scale with the materia resources already available in Pakistan. For this reason, the use of aircraft was rule out. Furthermore, the shower-clouds from which much of the monsoon rain falls ar generally short-lived and the efficacy of seeding was likely to depend on the salt particle being present in each cloud cell from the moment of its formation. This could only b achieved by seeding the atmosphere in general rather than individual cloud systems and could only be attempted by continuous dispersal from the ground.

2. Technique of salt-particle dispersal

(*a*) *Preparation and transport of salt*

According to the generally accepted theory of rain formation in non-freezing clouds an important part may be played by hydroscopic particles, whether natural or artificial containing between 10^{-9} and 10^{-8} g of salt. The lower limit is set by the necessity fo the particles to grow by condensation to a diameter of at least 40 μ within a few minute of entering cloud. The upper limit to particle size for dispersal from the ground is se by the velocity of fall, which increases rapidly for particles heavier than 10^{-8} g.

Method of dispersal of the salt particles.

The grinding of the salt was therefore carefully supervised. The crude salt (from he Pishin district in Baluchistan) was first spread out to dry in the sun for 2 to 3 days. ull advantage was taken here of the dryness of the air at Quetta, where the average elative humidity at midday in May and June is about 10 per cent. During the grinding f the salt, which was carried out by Messrs. Pak Flour Mills of Quetta, checks were made rom time to time of the particle size. This lay generally within the desired range, lthough it was not possible to avoid the presence of a small proportion of larger particles. ight tons of finely ground salt was prepared in this way.

After grinding, the salt powder was again laid out to dry for 24 hr before being oured into kerosene tins in which it was transported to the Punjab. The two holes in ach tin to which pipes were to be fitted for the dispersal of the salt (see following section) ere sealed by soldering metal discs over them. During the transport by goods train o Lahore, about 20 per cent of the tins were badly damaged, and for any future operations nore robust containers would have to be employed.

(*b*) *Dispersal of salt*

The method of dispersal of the salt particles is shown in the photograph (Plate X).

A hand-operated centrifugal blower of the type used by silversmiths in Pakistan orced air through the tin at high speed. Before entering the tin, the air passed through metal tube from which was slung a gauze basket containing burning charcoal. This aised the temperature of the air by 5-10°C and served two purposes : (i) to reduce the elative humidity of the air entering the tin and thus to prevent coagulation of the salt articles; (ii) to give the air leaving the tin some slight lift which helped to ensure that he particles were carried upwards from the ground.

Hand-operated blowers were preferred to power-driven blowers, since it was intended o try out a method which could be used anywhere regardless of the facilities available. All parts of the apparatus were of local manufacture, and the cost of one complete dispersal nit was approximately 75 Pakistan rupees (approximately £8 sterling).

It was found that the contents of one tin, 20 kg of salt, could be blown into the tmosphere in 30-40 min. This represents an output of salt particles, allowing a 50 per ent loss by fall to the ground in the immediate vicinity, of approximately 5×10^9 particles/sec.

On special occasions when the weather situation seemed particularly favourable, he salt was poured directly out of the tins from the top of a high building. In this way, he salt could be dispersed for short periods at a rate of 2×10^{10} particles/sec.

3. Seeding programme

Since the results of seeding were to be evaluated solely from the rainfall recorded by the network of raingauges already existing in the Punjab, the density of this network was an important factor in the selection of the sites from which the salt particles were to be dispersed. A second consideration was the natural variability of the rainfall.

These factors determined the choice of Lahore as the principal seeding station. It ies on the plains of the Punjab and there are no hills of any importance within 150 km, except to the north-east. The normal annual rainfall increases gradually from south-west to north-east across these plains, owing to the influence of the Himalayan foothills which lie some 200 km north-east of Lahore.

A second station was installed at Jauharabad in the Thal area. This was intended to seed the air converging toward the Salt Range, where orographic cumulus clouds are frequently observed. Jauharabad, 210 km WNW. of Lahore, could be assumed to be outside the range of influence of the seeding at Lahore.

There were no 'target areas' as such and, in order to limit the areas of action of the dispersed particles, seeding was done at the two stations only when the surface wind blew from an easterly direction. Furthermore it was limited to those days on which some low or unstable medium cloud was present. These two conditions are usually fulfilled simultaneously, since periods of easterly wind usually correspond to invasions of monsoon air masses. The cloud is mainly of convective type, with cloud base normally at 600-1,000 m and base temperatures above 20°C. Seeding was actually done at Lahore and Jauharabad on 39 days during the two months' period of the trials.

Since we relied on thermal convection to carry the particles aloft, seeding was done only during daylight hours and normally between 0800 and 1730 hr local time. On days of marked cumulus activity, up to 250 kg of salt was dispersed from each station.

4. Collection and analysis of rainfall data

By courtesy of the Director of Land Records and of the Executive Engineer, Irrigation Department, Lahore, special printed postcards were issued in advance to 186 provincial and irrigation department raingauge stations in central and northern Punjab. Each observer was requested to send one card daily to the field office at Regional Meteorological Centre in Lahore, stating the amount of rainfall recorded in the previous 24 hr and also the times of beginning and ending of rainfall.

This arrangement enabled the analysis of results to be made with the minimum of delay. The final evaluation, however, was made on the basis of the official rainfall data published by the Punjab Government agencies.

After classification and checking of the cards, maps were prepared showing the daily rainfall throughout the period 16 July-15 September over Central and Northern Punjab. The rainfall totals were plotted for this period as well as these totals expressed as percentages of the normal rainfall. To establish the normal rainfall distribution we had at our disposal 42-yr records from 67 stations in the area, and 14-yr records from a further 78 stations. Normals for the remaining stations were estimated by interpolation.

5. Evaluation of results

Samples of the natural hygroscopic particle content of the air in the Punjab, taken twice weekly during the previous 2 yr, indicated that during the monsoon season natural particles of mass greater than 10^{-9} g were likely to be present in concentrations of the order of 5 to 10 per m^3.

The processes of diffusion in the atmosphere on a large scale are not yet clearly understood, but a rough estimate shows that if particles are dispersed from a point source at the rate of 5×10^9 per sec in an average wind of 5 m/sec (normal for the Punjab), then their concentration downwind will fall to $10/m^3$ at a distance of approximately 100 km from the source (Sutton, *Micrometeorology*, 1953).

If the dispersal of the salt particles were to have any noticeable effect on the rainfall distribution, this should therefore be observed within 100 km downwind from the seeding station.

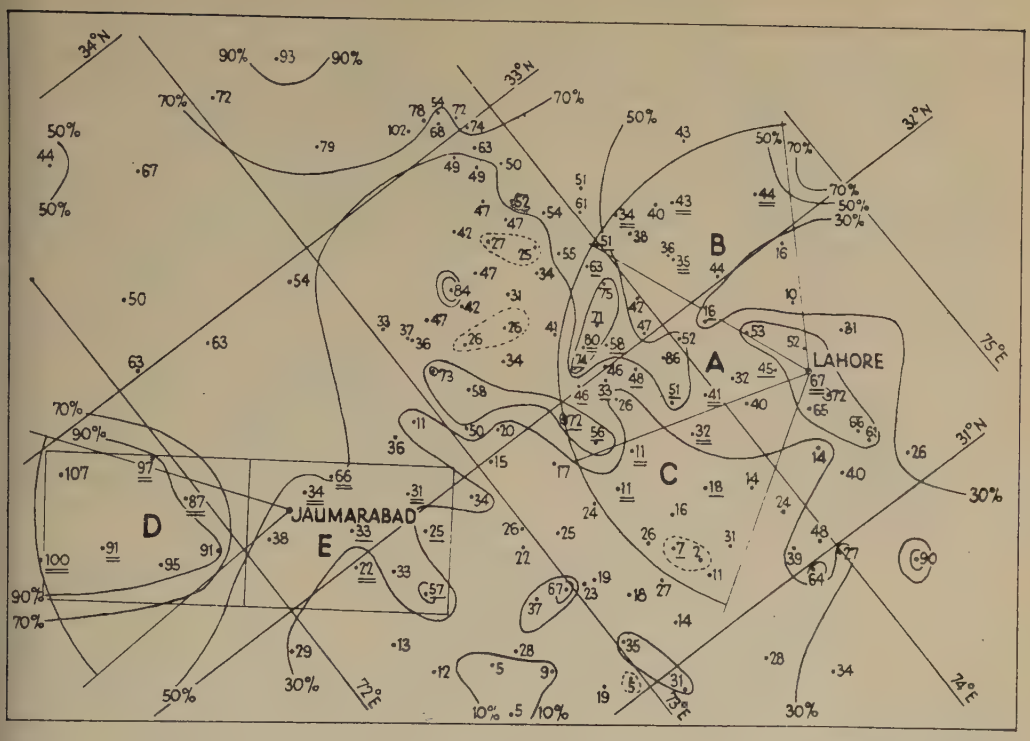

Figure 1. Rainfall in the Punjab for the period 16 July to 15 September 1954, expressed as percentages of July-August normals. Stations used in evaluation are underlined : *once* – 14 years past data; *twice* – 42 years past data.

On the accompanying chart (Fig. 1) the rainfall in the Punjab during the period 16 July to 15 September is shown in the form of percentages of the normal rainfall for the two months July and August. From the past records it was not possible to obtain normals for the period of seeding, but as the normal rainfall distribution pattern of September is very similar to that of July and August, this was not considered likely to be a serious source of error. The rainfall during the period was generally deficient over the Punjab, being less than 50 per cent of normal except in the submontane region to the north and north-east and in the Mianwali district. A number of stations near Lahore recorded relatively high rainfall, particularly to the north-west in the region of Hafizabad.

Seeding was done at Lahore on 39 days, during which the surface wind direction varied between NNE. and S. On 28 days (72 per cent of occasions), the wind direction lay between ESE. and SSE. With centre at Lahore, we have therefore drawn on the chart an arc of 100 km radius divided into three segments. Segment *A* is limited by radii drawn in the directions 290° and 340° from Lahore, thus enclosing 72 per cent of the wind directions during seeding. Segments *B* and *C* are limited by radii drawn from Lahore in the directions 030° and 240° respectively. For the purpose of evaluation, Lahore itself has been taken to lie in Segment *A*. Evaluation of the results of seeding at Lahore is based on a comparison of the rainfall received in Segment *A* with that in Segments *B* and *C* in 1954 and in previous years.

In the second area of seeding, the distribution of raingauge stations does not allow us to use similar segments. We have therefore drawn two contiguous rectangles, *D* and *E*, one of which lies downwind of the seeding station in Jauharabad and the other mainly upwind.

For the purposes of evaluation, areas *A* and *D* have been taken as the ' seeded areas,' and areas *B*, *C* and *E*, as ' control areas.'

During the two months' period of seeding in 1954, the average rainfall recorded at stations in the two ' seeded areas ' was in each case higher than that in the corresponding control areas. To judge the significance of this result, we have examined the past rainfall data to discover how frequently a similar rainfall pattern had been produced by natural causes.

Two series of past records were available : a 42-yr series of data from a small number of stations in each area; and a 14-yr series from a larger number of stations. The stations used are underlined on the chart (Fig. 1), twice for the long-period data, once for the shorter period. Table 1 shows the number of stations in each area for which past records are available.

TABLE 1. NUMBER OF RAINFALL STATIONS FOR WHICH RECORDS ARE AVAILABLE

Period	Area *A*	Area *B*	Area *C*	Area *D*	Area *E*
1901-1942	3	4	3	3	4
1928-1939 and 1941-1942	14	5	5	4	7

Only these same stations have been used in estimating the average rainfall in each area during the seeding period in 1954.

The average rainfall was calculated in each area for each year, by taking the arithmetic mean of the actual rainfalls recorded by the stations in the area during the months of July and August.

Thus, for the long-period records :

$$A = \tfrac{1}{3}(A_1 + A_2 + A_3)$$

$$B = \tfrac{1}{4}(B_1 + B_2 + B_3 + B_4)$$

where A_1, A_2, A_3, B_1, B_2 etc. are the actual rainfalls recorded at the individual stations during July and August each year.

The relations between average rainfall in neighbouring areas are then expressed as the logarithms of the ratios of average area rainfall. Thus :

$$\log_{10}(A/B) = \log_{10}\frac{\tfrac{1}{3}(A_1 + A_2 + A_3)}{\tfrac{1}{4}(B_1 + B_2 + B_3 + B_4)}$$

In this way we have derived values for each year of the following quantities :

$$\log(A/B),\ \log(A/C),\ \log(2A/(B + C)),\ \log(A/(BC)^{\frac{1}{2}}),\ \text{and}\ \log(D/E)$$

In Fig. 2, the values of these quantities in each year of the 42-yr period are shown in the form of frequency histograms, the years being grouped by intervals of 0·10 in the values of each quantity.

In 1954, the value of $\log(2A/(B + C))$ during the period of seeding was $+0{\cdot}22$, that is to say the average rainfall at the three stations in Area *A* was 66 per cent higher than the arithmetic mean of the average rainfalls in areas *B* and *C*. This value was only exceeded once between 1901 and 1942 (this was in 1929, when the value of $\log(2A/(B + C))$ was $+0{\cdot}23$).

It is interesting to note that the average rainfall in area *A* approaches the geometric mean of the rainfalls in the neighbouring areas more closely than it does their arithmetic mean. In 1954, the value of $\log(A/(BC)^{\frac{1}{2}})$ was $+0{\cdot}31$, that is to say that the three stations in area *A* received more than double the geometric mean of the average rainfalls in areas *B* and *C*, a ratio which was never exceeded in the period 1901 to 1942.

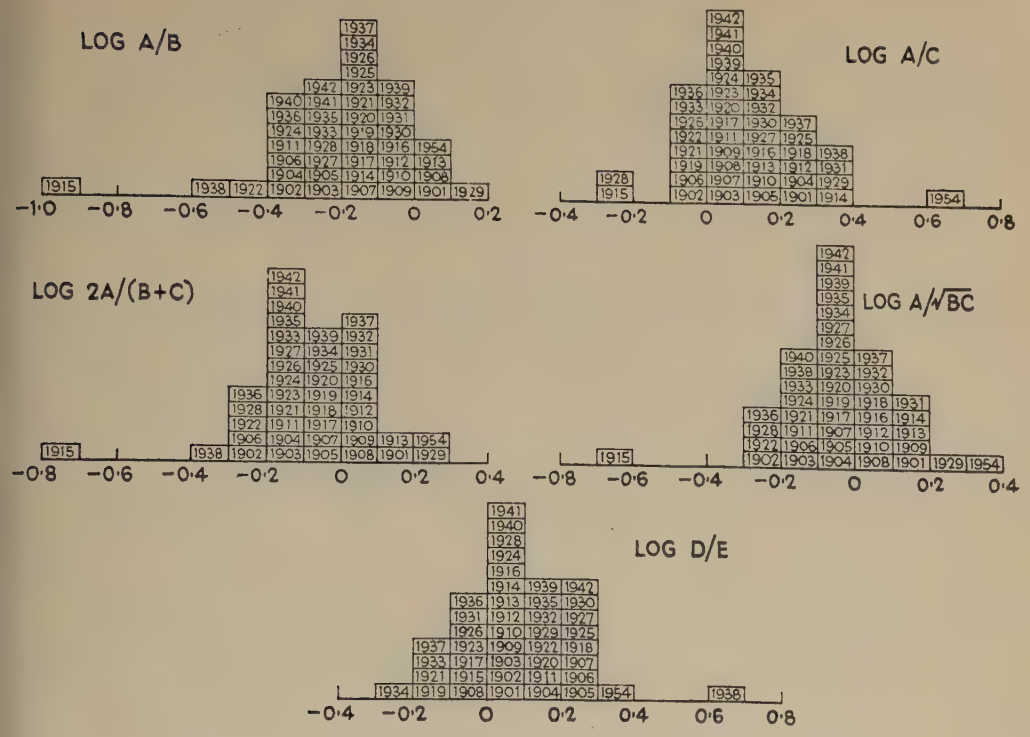

Figure 2. Frequency histograms of comparative rainfalls (42-yr period).

Turning to the second seeding station, we see the value of log (D/E) in 1945 was + 0·38, which was exceeded only once before, in 1938.

The number of stations for which 42-yr records are available is rather small, and we supplemented this series by a second overlapping series from 1928-1939 and 1941-42, for which data for a larger number of stations are available. The number of years in this case was too small to construct frequency histograms, and so the individual values for each year are given in Table 2.

TABLE 2. 14-YR SERIES (average rainfall in areas A-E given in in.)

Area	1928	1929	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939	1941	1942	Mean	1954
A	5·22	13·16	13·11	12·37	11·46	18·44	12·25	9·12	5·71	8·20	5·43	5·76	7·42	11·36	9·93	5·26
B	10·35	12·81	13·49	18·82	10·46	26·22	17·07	18·08	12·23	11·79	13·98	7·08	10·48	16·68	14·25	6·04
C	7·74	6·85	8·11	8·86	6·01	14·64	6·50	4·45	6·44	4·65	2·00	4·88	6·91	7·31	6·81	1·35
D	6·43	15·23	4·73	3·60	12·55	10·49	5·53	7·75	4·77	3·71	4·11	5·43	7·43	7·61	7·10	5·95
E	4·81	10·90	4·89	4·70	8·45	14·20	10·79	6·44	6·31	5·40	1·04	4·56	8·12	5·19	6·84	2·50
$\log(A/B)$	−·30	+·01	−·01	−·18	+·04	−·15	−·14	−·30	−·33	−·16	−·41	−·09	−·15	−·17	−·16	−·06
$\log(A/C)$	−·17	+·28	+·21	+·14	+·28	+·10	+·28	+·31	−·05	+·25	+·43	+·07	+·03	+·19	+·16	+·59
$\log\frac{2A}{B+C}$	−·24	+·13	+·08	−·05	+·14	−·05	+·02	−·09	−·21	−·00	+·17	−·02	−·07	−·02	−·03	+·15
$\log\frac{A}{(BC)^{\frac{1}{2}}}$	−·24	+·15	+·10	−·02	+·16	−·03	+·07	+·01	−·19	+·05	+·01	−·01	−·06	+·01	0·0	+·26
$\log(D/E)$	+·13	+·15	−·02	−·12	+·17	−·13	−·29	+·08	−·12	−·16	+·60	+·08	−·04	+·17	+·02	+·38

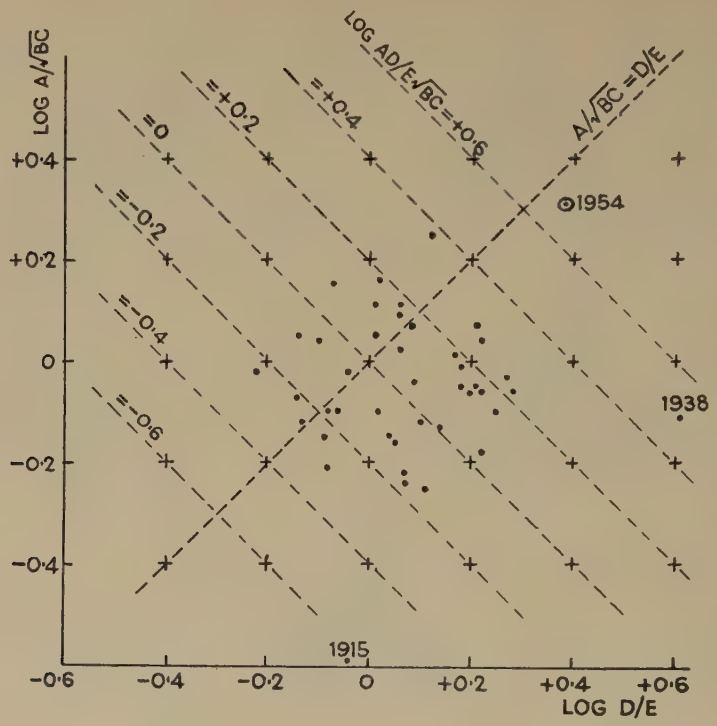

Figure 3. Relation of log $A/(BC)^{\frac{1}{2}}$ to log D/E (42-yr period).

The values of the five quantities in this series are not always equal to those determined for the same year in the 42-yr series, since more stations have been used in their determination, but the final result remains unchanged. The new value of log $(2A/(B + C))$ for 1954 is + 0·15, which was approached closely in 1929 and 1932, but never exceeded in this 14-yr period. The value of log $(A/(BC)^{\frac{1}{2}})$ for 1954, + 0·26, is far in excess of any previous value. The value of log (D/E) for 1954 is still + 0·38, which was only approached and exceeded once, in 1938.

There was thus in both the ' seeded areas ' significantly higher rainfall in 1954 than in the ' control areas.' This is brought out in Fig. 3, in which the values of log (D/E) for each year of the 42-yr period have been plotted against the corresponding values of log $(A/(BC)^{\frac{1}{2}})$. The 1954 values of log $(A/(BC)^{\frac{1}{2}})$ and log (D/E) differ in each case from the mean values of these quantities by more than twice their standard deviations. The value of their sum, log $(AD/E(BC)^{\frac{1}{2}})$, in 1954 was the highest ever recorded. Furthermore, the 1954 values lie close to the line of equal values of the two quantities, which would be the case if the rainfall in both areas A and D were influenced to the same extent by the seeding.

6. Discussion of results

The above analysis shows that the distribution of rainfall in the central Punjab during the period of seeding in 1954 was distinctly abnormal. The problem is whether we can be sure this was actually due to the seeding.

In the absence of direct confirmation that the dispersed salt particles actually entered the rainclouds in the ' seeded areas,' we cannot yet answer this question by a definite affirmative. It does appear, however, that the abnormally high rainfall did occur in those areas to which the dispersed salt particles were most probably carried by the prevailing wind.

To detect and count the dispersed salt particles in the air downwind of the seeding stations would require aircraft and equipment beyond our present or probable future resources, but it is intended to repeat these trials at the same stations and with the same technique during the 1955 monsoon season, when a repetition of the same abnormalities in rainfall distribution, if observed, would be regarded as justifying further experiments on a larger scale.

7. Acknowledgments

We wish to acknowledge the valuable assistance accorded to us by the Director of Land Records, Lahore, and by the Punjab Irrigation Department in arranging the special reporting of observations from their networks of raingauge stations.

One of us (E.M.F.d'A.) was in Pakistan as member of a UNESCO Technical Assistance Mission.

Mr. Quayum Anwar and Mr. Azhar Adil, of the Regional Meteorological Centre at Lahore, greatly helped in the local arrangements for the trials.

Much of the burden of the analysis of rainfall observations was borne by Mr. Mohammad Shaffi Ahmed and Mr. Abdul Halim, of the Geophysical Observatory, Quetta.

Finally, we wish to record our thanks to Mr. S. N. Naqvi, Director of the Pakistan Meteorological Service, for his valued support and encouragement in the planning and conduct of these experiments.

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Methods of synoptic analysis at low latitudes

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SUMMARY

The apparent differences in the methods of synoptic analysis employed in the tropical Pacific and in equatorial SE. Asia represent nothing more than adaptations of the same technique. Over the Pacific it is unusual for two different air-streams to meet at a large angle and, since their common boundary cannot be traced from day to day with continuity, reliance must be placed on recognition of short-lived asymptotes and vortices from the instantaneous streamlines. Over equatorial SE. Asia, long semi-permanent boundaries can be traced with continuity owing to the persistence of the westerlies. The difference of opinion as to the most desirable method of analysis is mainly due to the variations in horizontal extent of the equatorial westerlies in different longitudes. At the boundary between the trades and the westerlies there are separate centres of rainfall rather than a continuous rain-band.

1. THE TWO METHODS OF ANALYSIS

Although weather analysis for equatorial regions now appears superficially to be developing along two distinctly different lines, these are really only adaptations of the same technique according to locality. In equatorial SE. Asia, the most important features of the streamline charts are the boundaries between the equatorial westerlies and the trades from either hemisphere, because convergence is common along these boundaries (Watts 1949). In recent examples of this type of analysis afforded by Schmidt (1951) and Thompson (1951), long semi-permanent boundaries are drawn between different streams of air. Watts (1955a) calls these discontinuities 'air-stream boundaries' to avoid the criticism made by Palmer (1952a) on the use of the expression 'front.' The introduction of the new expression is desirable because, although temperature-contrasts between different air streams (Sawyer 1947) may exist over the continents, there is no evidence of such contrasts over the equatorial oceans.

The methods used in SE. Asia, in which continuity is so important, appear to be inapplicable in the tropical or equatorial Pacific. Palmer (1951) considers that the trades in the Pacific do not meet along a continuous line, as previously visualized by Brooks and Braby (1921), but in a series of short-lived asymptotes and vortices. Palmer (1952b) further states that, in the western Pacific (east of equatorial SE. Asia), any extensive bands of equatorial westerlies result from a train of vortices closely following one another. It is thus implied that, although the westerlies may appear on mean maps, they are not a permanent feature of the general circulation.

Here we have the cause of the difference in analytic method. Between 10°N and 10°S, and west of 160°E, it is claimed (Watts 1955a) that the westerlies and one or two other streams usually meet at a large angle and that the semi-permanent boundary between the streams may be traced with continuity from day to day. East of about 160°E the westerlies and the asymptotes of convergence are short-lived, so that continuity cannot be achieved for long periods. It also appears that the style of streamline analysis used by Schmidt (1951) and others is not readily applicable to the problems of the western Pacific, where the isogon analysis of Bjerknes, Hesselberg and Devik (1911) is favoured as more clearly defining the short-lived vortices; but, as Palmer (1952a) points out, there is no qualitative difference in technique.

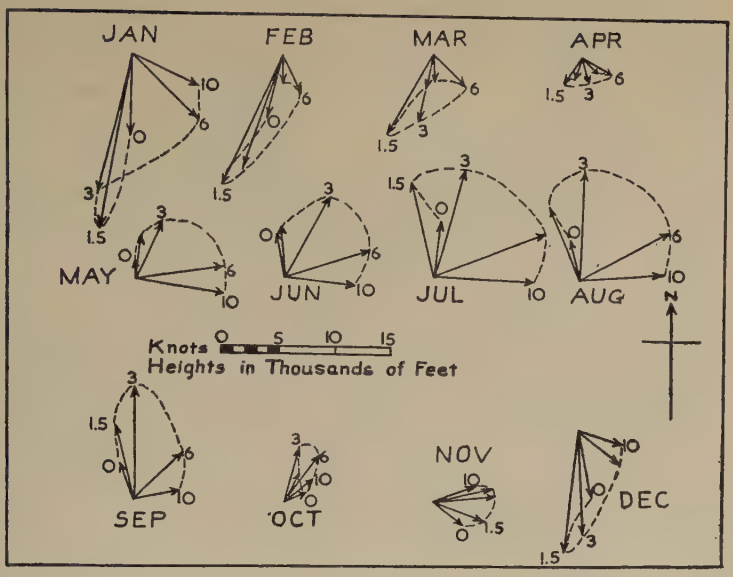


Figure 1. Vector mean winds over Singapore, 0620-0820 LMT (8 years' observations 1934-41).

The validity of the continuity method depends on whether or not the westerlies are persistent. Although Palmer's (1952b) claim that the westerlies are due to a series of vortices following one another may be true for the Marshall Islands area, it does not seem to hold for the region further westward, where westerlies are present between 10°N and 10°S on nearly every day of the year. Thus, vector mean winds at Singapore (1°N) show marked westerlies between 6,000 and 10,000 ft from November to January and from May to September (Fig. 1), while Djakarta (6°S) has a deep layer of westerlies throughout January and most of December, February and March. The significance of the vector mean winds over Singapore is evident from Fig. 2, which shows that the west-to-east component was positive on 102 out of 122 days during June, August, September and October 1952.

2. Origin of the westerlies

Since the westerlies are most pronounced in the summer hemisphere and in longitudes where monsoons are strongest, there is some reason for considering that they may derive from the same cause (i.e. differential heating of the continents and oceans). This cannot be the sole cause, however, because the westerlies can frequently be traced back as a continuous stream far to the west of the Indian Ocean, and they often cover Indonesia when no monsoon effect can be detected over Australia.

Fletcher (1945) explains that there are two latitudinal axes of convergence, one on either side of the equator, and that the air at low levels between them is a cold source because of radiation from the cloud-tops. Then, under geostrophic control, the wind on the poleward side of each convergence-line is an easterly, while westerlies develop on the equatorward side. The disadvantages of this theory are that (a) it is improbable that geostrophic control is effective at low latitudes, (b) there is no evidence that the fall of temperature through radiation from the cloud tops would sufficiently exceed the adiabatic warming of the subsiding air to promote a cold source at low levels, and (c) the westerlies are most often confined to the summer hemisphere rather than being in a broad stream extending over both sides of the equator.

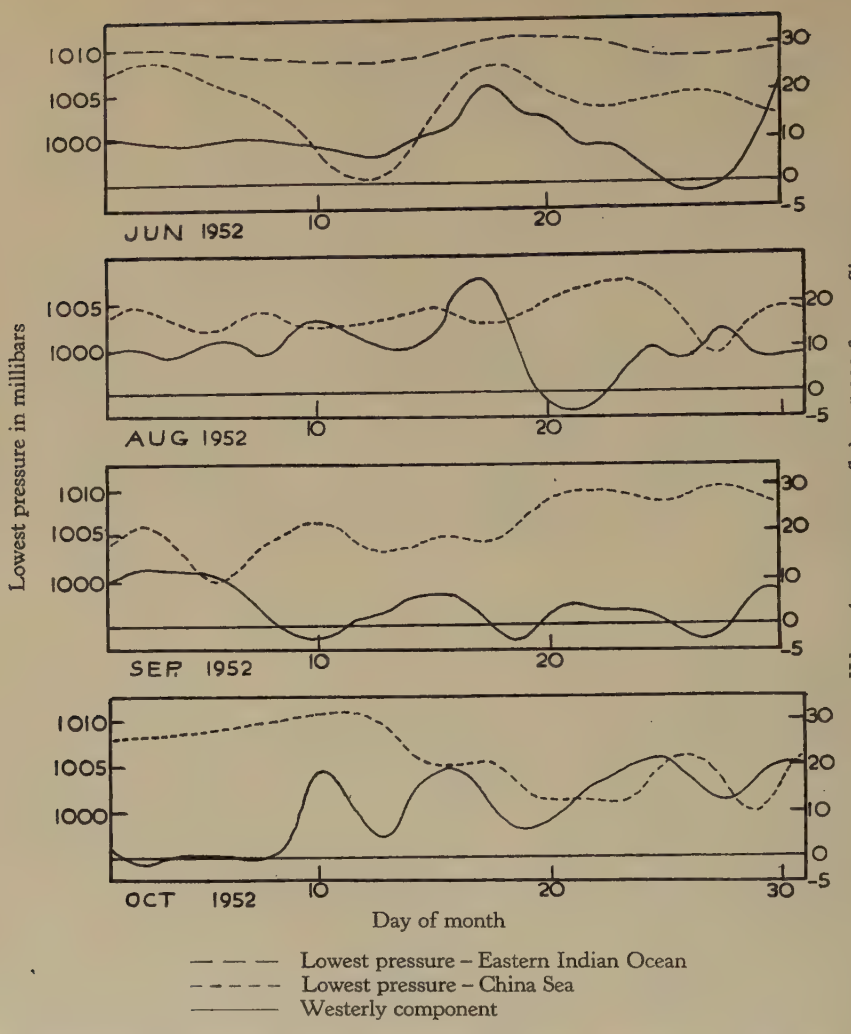


Figure 2. Comparison of Singapore westerly component and lowest pressure in the China Sea and eastern Indian Ocean.

The westerlies of equatorial SE. Asia are usually traced on the daily streamline charts as consisting of air which has crossed the equator, and this effect may be seen (Fig. 3) from the mean distribution in July of winds at 5,000 ft over Medan (03°34′N), Singapore (01°18′N), Palembang (02°54′S) and Djakarta (06°09′S). Similarly, the daily charts for the southern summer show that the westerlies over Java usually consist of air which originally crossed Malaya in north-easterlies. It might be considered that the air crossing the equator in the northern summer is drawn in to feed the SW. Monsoon over the Indian Ocean. Individual streamline charts show that the flow is initially from the south-east along the isobars of the southern hemisphere. North of the equator, isobars curve away to the north-west, with lowest pressure lying to the north-east. Now this is one of the cases proposed by Grimes (1937), who shows that, with such conditions, the wind would turn to a south-westerly flow north of the equator to cross the northern-hemisphere isobars at a large angle. As Crossley (1946) points out, Grimes makes certain assumptions which detract from the value of the solution (namely, horizontal motion independent of longitude, steady flow, constant density and no frictional forces). Despite these deficiencies,

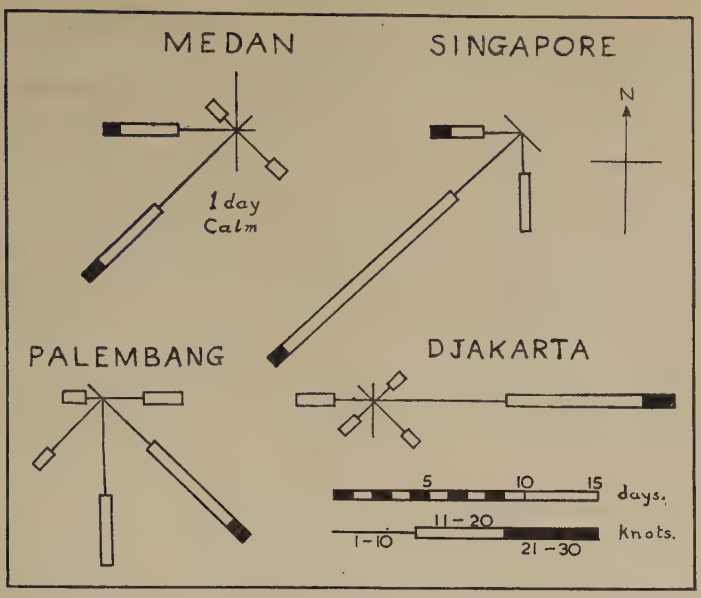


Figure 3. Mean distribution of winds at 5,000 ft in July (4 years' observations 1936-39).

he solution does fit the observed facts. It does not preclude the possibility that the trength of the south-westerlies may vary to some extent with the frequency of typhoons n the far north-east.

Grimes's solutions do not explain how the NE. Monsoon, which initially crosses the sobars at a large angle in the South China Sea, turns to westerlies south of the equator; ut here Grimes's patterns are difficult to apply, since pressure gradients just south of he equator are very small and the isobaric pattern indistinct. Schmidt (1951) has explained ow westerlies should result south of the equator in this season provided there is a line f convergence in the south, and Watts (1949) has shown that the latter condition does isually exist from November to March.

Schmidt (1951) points out that the westerlies are not necessarily confined to streams vhich have crossed the equator, but that they sometimes lie in a broad stream on both ides of it. The occasional development of westerlies in the winter hemisphere could lepend on the presence of cyclonic vortices as Palmer (1952b) suggests; but this is difficult o determine because, although cyclones may not always be present, one or more shallow ow-pressure areas usually exist somewhere in the region without sufficient upper-wind bservations to define the circulation around them.

3. Relation of the westerlies to cyclone series

To examine whether the strength of the westerlies at 5,000 ft over Singapore varies vith the incidence of cyclones in either hemisphere, the westerly component of the vind at 00 GMT on each day of each month has been plotted together with (a) lowest ressure on each day in the China Sea between 8°N and 20°N and between 110°E and 22°E (averaged from the 00 and 06 GMT synoptic charts) and (b) the lowest pressure n each day in the eastern Indian Ocean between the equator and 15°S, 95°E and 125°E. The 5,000-ft wind was chosen in these comparisons because this is the lowest level over ingapore which is free from orographic sheltering by the Sumatran ranges.

It was found that deep lows were rare in the eastern Indian Ocean at any time of he year, so it is unlikely that they contributed much to the westerlies. Storms in the

region east of the Philippines were not included in the analyses as they were considered too remote to influence directly the Singapore winds.

In May 1952 only one short-lived tropical depression occurred in the eastern Indian Ocean and none in the China Sea or west Pacific. Despite this, the Singapore 5,000-ft wind had a westerly component on 19 days and an easterly component on only 12 days. The easterly components were all less than 7 kt but the westerlies reached 18 kt.

The lowest pressures and the westerly components during June, August, September and October 1952 are shown in Fig. 2. A relation might be suspected between the strength of the westerlies and the incidence of typhoons in the China Sea during August and from 20 September to 10 October, but no such relation is evident in other months. In October 1952 there were more than twice the normal number of typhoons in the China Sea according to the Hong Kong Meteorological Service (1952), yet the Singapore westerlies in that month were not stronger than in the other months. For instance, October 1952 had 15 days with the Singapore westerly exceeding 10 kt, whereas November 1952 had 24 such days. On the other hand, westerlies between 5,000 and 10,000 ft over Singapore were more frequent and stronger in October 1952 than is normal for October.

Westerlies were strong throughout November 1952, although only two marked centres of low pressure occurred in the China Sea and none in the Indian Ocean.

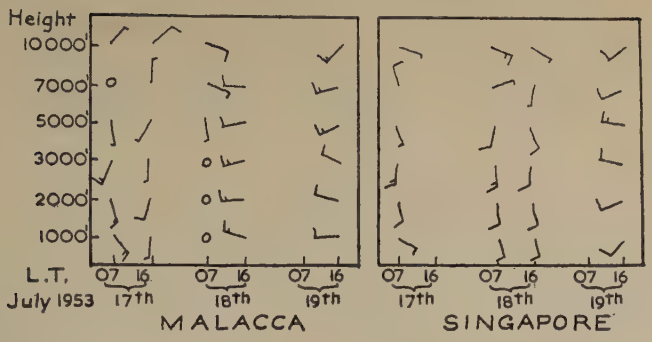
It appears, therefore, that although the strength of the Singapore westerlies may depend to some extent on typhoons in the China Sea (but not south of the equator), the westerlies also persist in the absence of any discernible vortices and must be considered as a major feature of the general circulation. In fact it is quite possible that typhoon formation in the China Sea *follows* a strengthening of the westerlies near the equator, since the latter would contribute to the degree of wind discontinuity between the westerlies and the trades, which Thompson (1951) considers a prerequisite for typhoon formation.

4. Longitudinal extent of the wind discontinuity

Over the land masses of Sumatra, Malaya, Borneo and Java, where plentiful observations enable recognition of the westerlies as a permanent feature, the wind discontinuity usually appears to be along a line which can be traced in streamline analyses from one hour to another and from level to level without undue difficulty. It is more difficult to determine how far a continuous boundary between westerlies and trades may extend to east and west of this region.

Over the eastern Indian Ocean and the seas east of Java, and between 12°S and 10°N, evidence of a meeting between westerlies and south-easterlies is often found from aircraft and ships' reports; but such reports are not frequent enough at each synoptic hour to prove that the westerlies and the air-stream boundary are permanent features there. However, there is every indication from the mean surface-wind roses (Meteorological Office 1945, 1952) that the boundary is a permanent feature near the equator as far eastward as 115°E. These roses show that, from June to September, westerlies from the Indian Ocean between 5°N and 10°N extend eastward to 115-120°E. Similarly, those between 5°S and 10°S during December to March extend to 125-140°E.

The wind roses also show that westerlies predominate near the equator, at least as far westward as 85°E, throughout the year. From May to October they extend well northwards as part of the Indian SW. Monsoon and westward to 50°-70°E. During December to March they lie between the equator and 5°S and extend westward to 70°-85°E; during April and November they still predominate as far westward as 60°E, though in a narrow zone near the equator. Thus there is every indication that semi-permanent discontinuities should be found over at least 30° of longitude.



gure 4. Upper winds over Malacca and Singapore showing a change of air-stream (short feather 5 kt; long feather 10 kt).

5. Rainfall at an equatorial air-stream boundary

Although the boundary between the westerlies and the trades is such an important ature of the Asian analysis it cannot be assumed that there is a continuous belt of rainfall the boundary. For example the equatorial westerlies replaced the SW. Pacific trades ver Singapore between 16 LMT 18 July and 16 LMT 19 July 1953 (Fig. 4). That the ind-change was not purely local is evident from similar changes at Malacca (Fig. 4) and ort Swettenham. Over Singapore Island a close network of 30 recording raingauges nowed that there were separate centres of light to heavy rainfall from 20 to 24 LMT 8 July and from 06 to 0915 LMT 19 July. At other times on both days, rain was negligible. broken rain pattern like this was associated with nearly every change from southerlies westerlies during 1953, throughout which year the close network was maintained. On ne occasion, one gauge recorded 4·66 in. of rain in an hour, and yet the rain band ssociated with the air-stream boundary still consisted of a series of individual storms Watts 1955b).

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An operational frontal contour-analysis model

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SUMMARY

A three-front model is proposed as a practical solution to analysis problems associated with both surfa fronts and the baroclinic currents of the upper westerlies. They are called the ' polar front,' the ' maritir front,' and the ' arctic front.'

A front is defined as a three-dimensional baroclinic zone, having a first-order discontinuity in t temperature and wind fields, and being reasonably continuous in space and time. Frontal contour analy is found to be operationally practicable and desirable.

1. INTRODUCTION

For the past three years a group of meteorologists assigned to the Central Analys Office has been developing operational techniques for representing the three-dimensionr structure of the atmosphere. It became apparent that a method was required to effe analytical integration of the atmospheric layers to which attention is normally directed.

The lower layer is traditionally analysed by the surface (synoptic) chart whic continues to be the major tool at most operational weather offices. The discrepanci revealed by widespread dissemination of frontal analyses over the past decade indica that even a detailed examination of surface hourly and synoptic reports is not sufficier to provide a unique analysis that will satisfy historical continuity. The definition ar organization of the surface frontal systems presented an acute problem.

The intermediate layer embraces the 700 and 500-mb charts where many surfa complexities disappear and the flow assumes a more uniform and symmetric patter Among operational staffs the so-called steering properties of this layer have found gener use. The broad pseudo-barotropic features have permitted more objective analysis ar prognostic techniques. The important analytical problem lies in establishing the structur of the thermal field in order to differentiate between (essentially) baroclinic and barotrop features.

The upper layer includes the 300 and 200-mb charts and features the strongl baroclinic properties of the jet streams. The structure of the flow in this layer cannot b obtained independently, except for limited areas where rawin data are plentiful.

Thus the individual analysis problems are, in the main, associated with the baroclin properties of the atmosphere. It has been found that the frontal contour chart not onl provides an adequate solution to the problem of surface frontal analysis but als organizes the major baroclinic zones so that a reasonable structure of the jet streams ma be determined even in areas where rawin data are quite limited.

This paper presents the major features of a system for three-dimensional front analysis, the detailed implications being reserved for separate treatment.

2. THE ANALYSIS TOOLS

Six-hourly surface and twelve-hourly 700-mb, 500-mb, 300-mb and frontal contou charts were used. The frontal contour chart was based primarily on tephigram plo to 100 mb of all radiosonde data and shear hodograph plots of all rawin data; this analys

ıkes 1½ to 2 hr. In addition, routine vertical cross-sections were analysed and a variety f special charts tested. Detailed analyses covered an area extending longitudinally from hip Sierra in the Pacific to ship Coca in mid-Atlantic; the surface-pressure and upper ontour fields were extended eastward to Central Europe.

3. Frontal concepts

Clarification of frontal concepts became necessary early in the project. Literature n fronts reveals that general agreement is lacking in this regard and the deficiency has een dealt with in some detail by Sutcliffe (1952). Palmén (1951) describes a front as a loping layer of maximum horizontal temperature gradients with a first-order discontinuity ı the wind field. Godson (1951) used the term ' hyperbaroclinic zone ' for a frontal ayer with a first-order discontinuity in the temperature field. These two concepts are hysically and mathematically identical and appear to be realistic definitions of the con- itions that prevail at fronts. However, they are not sufficient to define a front uniquely. imilar conditions are associated with certain types of dry inversions and marine strata. o eliminate these phenomena and to allow successful charting on the basis of current pper-air networks the following frontal definitions were adopted :

(i) It is a three-dimensional hyperbaroclinic zone with a first-order discontinuity ı the temperature and wind fields. Mathematical criteria have been developed on this asis.

(ii) It is a quasi-substantial surface which moves with the wind flow.

(iii) It is a reasonably continuous feature of the chart both in space and time.

4. Frontal characteristics

The following features regarding fronts, some of which have been emphasized by arlier writers, became apparent from this study.

Three-dimensional fronts are a major dynamic feature of the extra-tropical tropo- phere.

The most suitable parameters for representing the frontal discontinuities on plotted harts were found to be :

(a) The orientation as given by the shear hodograph and (b) the wet-bulb potential emperature, θ_w, and the saturated wet-bulb potential temperature, θ_w', as given by the ephigram. θ_w is one of the more conservative atmospheric parameters and θ_w', repre- enting the pseudo-adiabat through the dry-bulb temperature, is perhaps the most epresentative value for characteristic temperature curves above frontal surfaces. In ddition, the pressure and potential temperature at the tropopause were found to have onsiderable value in frontal identification.

The top of the frontal zone is taken as the frontal surface since this point can be nore readily and consistently identified than the base of the frontal zone (Bjerknes 1930). Throughout the project it was found advisable to associate each frontal system, for both dentification and continuity purposes, with the warm air lying above the surface. Although his departs from the conventional system on frontal nomenclature, it does agree with he generally accepted surface-analysis procedure of locating a front at the edge of the elatively homogeneous warm air mass. Further it is the writers' opinion that confusion n this point is partly the cause for anomalous structures and historical discontinuity in nany surface frontal analyses, i.e. continuity is not maintained with respect to the warm ir. In addition, continuity of a jet-stream analysis in association with three-dimensional ronts becomes possible when fronts are maintained on the foregoing basis.

Routine analyses over the past two years have supported the conclusion that thr frontal systems are sufficient to delineate the major baroclinic zones of the atmosphe and that no more than three frontal systems can be expected to have significant dynam properties. A *three-front model* has thus been adopted as the most realistic solution the problems of 3-dimensional frontal analysis.

5. Frontal and air-mass nomenclature

A colour scheme was used for frontal display and reference to avoid a conditione approach based on a conventional air-mass classification. The three-front model hh thus evolved from an empirical study of frontal contour charts in conjunction with th dynamical aspects of the zonal circulation and does not require a classical air-ma approach. However, the two are reconcilable and operationally it is convenient to ado a nomenclature consistent with the results of the investigation and with precedence. Th use of the terms ' polar ' for the southernmost and ' arctic ' for the northernmost fro have become well established in meteorological literature. These terms are derive from an association of the front with the air mass to the north which is not in agreeme with the project findings. However, the historical names can be accepted providing stri adherence to this association is not presumed. The middle front does not have a simil basis in precedence. Two names are suggested : ' inter-polar ' since the front separat air masses of polar origin and ' maritime ' since the two air masses occur mainly throug maritime modifications. The latter name is preferred and will be used in this paper.

The classical air masses delineated by the three-front model may be grouped und the following general headings :

South of the polar front – tropical air; south of the maritime front – maritime pol air; south of the arctic front – maritime arctic air; north of the arctic front – continent arctic air.

Some readers may wish to identify maritime arctic in the above classification wit cold maritime polar air. However, in a classification based on the major fronts a broade use of the term maritime arctic is preferred.

6. The three-front model and the zonal circulation

Adoption of the three-front model has been based to a considerable extent on th role these fronts play in the day-to-day structure of the westerlies. The relationship i implied when a front is defined as a sloping layer of maximum horizontal temperatur gradients. The relationship with respect to two of the more frequent types of zona flow in the intermediate layer of the atmosphere may be summarized broadly as follows.

(*a*) *A single westerly stream*

The southern half of the stream is usually associated with the baroclinicity of th polar front. At 500 mb the polar front is found near the middle of the stream. Th northern section of the stream has a similar association with the maritime front. Th arctic front is usually very shallow above 850 mb and at 500 mb is related to the westerl stream only in its more intense sections.

Over considerable portions of the chart two jet streams, associated with the pola and maritime fronts respectively, are found. In the developmental areas associated wit major troughs the two fronts are found in close proximity at all levels and a single powerfu jet is normally observed.

Perturbations on the various fronts are usually in phase and a single system o depressions is normally found on the surface chart.

(*b*) *Split westerlies*

Although split westerlies seldom cover areas approaching hemispheric size they are uite common over areas covering up to 120 degrees of longitude.

In the southern stream the polar front is usually found well within the stream at 00 mb, whereas the maritime front is usually discontinuous in this branch, often taking he form of cut-off domes.

In the northern stream the maritime front at 500 mb is fairly continuous and lies ell within the stream. The arctic front occupies a more northerly position in the lower evels and is relatively steep. As a consequence it is actively associated at 500 mb with he northern branch of the westerlies.

Two well-defined jet streams are found in this case, one associated with the northern tream and the maritime front (reinforced by the arctic front) and one with the southern tream and the polar front.

In some cases separation of the maritime and arctic fronts may be associated with he appearance of two jet streams over the northern branch. Three jet streams may then e present at certain longitudes on the chart.

Perturbations on the various fronts are usually not in phase and two cyclone systems nay be found on the surface chart.

7. The three fronts and associated air masses

The discussion is confined to the winter season when characteristics are most clearly lefined.

(*a*) *The polar front*

The air mass above and to the south of the polar front is, in the main, the classical nT. Continental tropical is not an important winter air mass in the analysis area and or frontal purposes is always grouped with mT. The mT air and the polar front show onsiderable variations in different sections of the chart.

Over the Aleutians mT air is usually present at 500 mb and above during the levelopment of a major cyclone. With prolonged ascent the air will be nearly saturated vith values of θ_w generally as low as 12°-14°C due to radiational cooling. Values as igh as 18°C may occur in a strong meridional flow. The polar front is well marked n rawinsonde plots.

In the western USA, mT air of Pacific origin is usually present aloft and often extends down to the 800-mb level. Extensive subsidence on the east side of the Pacific high haracterizes this air, with dryness being further accentuated by the prevalent low-level ubsidence in the lee of the Rockies. θ_w values will range from 14°-16°C with θ_w' values eing more variable often reaching 20°C in the lower levels. The front may be quite lat, at times lying only a few thousand feet off the ground over a wide region; surface nalysis is then very difficult unless carried out in conjunction with a frontal contour hart.

In the south-eastern USA, mT air displays typical source-region properties having uniform moisture properties in the lower levels and wide variations aloft. When this air scends into the north-eastern USA and south-eastern Canada it becomes relatively omogeneous and is the main precipitation source for the region. In a zonal flow Pacific mT may be present aloft in the northern area and considerable baroclinicity is present n the mT air mass. In the Florida area θ_w values may be near 20°C at all levels but a decrease to about 15°C at 700-600 mb is common. Farther north values will vary from 13°-18°C depending on the strength of the zonal flow.

The polar front over North America in winter has average positions at about 25° on the surface and 38°N at 500 mb. The surface front seldom penetrates much beyon 45°N but the 500-mb position may extend to near 70°N in the western sections. Cut-c pockets of mT at the 500-mb level are not uncommon in northern latitudes but owing the strong radiational cooling they seldom last more than two days.

(*b*) *The maritime front*

Perturbations on the maritime front are closely associated with the formation an development of most frontal cyclones. Since these perturbations usually lie within th zonal westerlies their motions are fairly uniform and provide a valuable prognostic tool.

Prolonged trajectories in the westerlies over the Pacific Ocean provide the main sourc for mP air which affects North America. In some cases long-term radiation may transfor tropical air into mP.

Lapse rates for the mP approach the pseudo-adiabatic with θ_w values ranging fro 7° to 12°C at all levels. In a weak flow around the periphery of the Pacific high, di subsidence layers in the mP are common at intermediate levels. Over the cool coast waters of California the low-level mP becomes quite stable. Over northern USA an southern Canada the mP air usually does not reach the ground owing to the presence snow-cover. It is common then for troughs of mP air to move aloft across that regio without any mP evident at the surface. These troughs are usually associated with th development and motion of waves on the arctic front.

Although frontolysis and frontogenesis occur frequently in the lower levels the fro exhibits a high degree of three-dimensional continuity. At 500 mb it may, for mo practical purposes, be considered continuous in space and time.

The average surface position is about 40°N over the Pacific and off the Atlanti seaboard and lies south of the snow-line over the continent. At 500 mb the position about 55°N but in extreme cases the front may move well to the north of the Arcti Circle.

(*c*) *The arctic front*

The arctic front is the northernmost one of importance in the development of extra tropical cyclones. The mA air to the south is usually the coldest air which has sufficien moisture and instability to produce precipitation through ascent associated with fronta systems. A supply of the cold continental, cA, air found to the rear of the front is normall required for the development of one of the major winter cyclones.

The main source for mA affecting North America is the Pacific Ocean, north c latitude 45°. The convective transformation of cA over the water surface usually occur in cyclonic flow and the process is effective throughout the troposphere. The θ_w of th resulting mA air varies about a mean of 1°C.

In a strong northerly flow a well-marked arctic front can be located at *Weather Shi P* but in all cases studied frontolysis was completed to the west of the British Columbi coast. A similar process takes place in the north-west Atlantic. Despite the periodi frontolysis and frontogenesis in maritime areas the analysis of the front is usuall simple.

During the southward surge over the continent the frontal structure becomes quit complex. Over the northern USA and Canada snow-cover precludes strong surfac heating and south of an arctic high the surface front is well marked. Turbulent mixin in strong pressure gradients associated with cyclonic flow operates to destroy the charac teristic cA inversion, and south of the snow-line surface heating accomplishes the sam

result. The surface position of the arctic front then may be indeterminate over large areas of the chart. However, the southward intrusions of cA air are usually quite shallow and at higher levels the front remains well defined in a more northerly position.

(*d*) *The tropopause*

The day-to-day variations in the height and potential temperature of the tropopause associated with a particular front are usually quite small and thus serve as useful air-mass identifiers. Significant changes occur over long periods or near strong jet maxima. During the winter months the following values are fairly characteristic.

Air mass	Tropopause Height (mb)	Tropopause Potential temperature (°K)	Stratospheric mean temperature (°C)
cA	400	300	— 48
mA	300	310	— 52
mP	250	320	— 57
mT	200	330	— 62

8. Examples

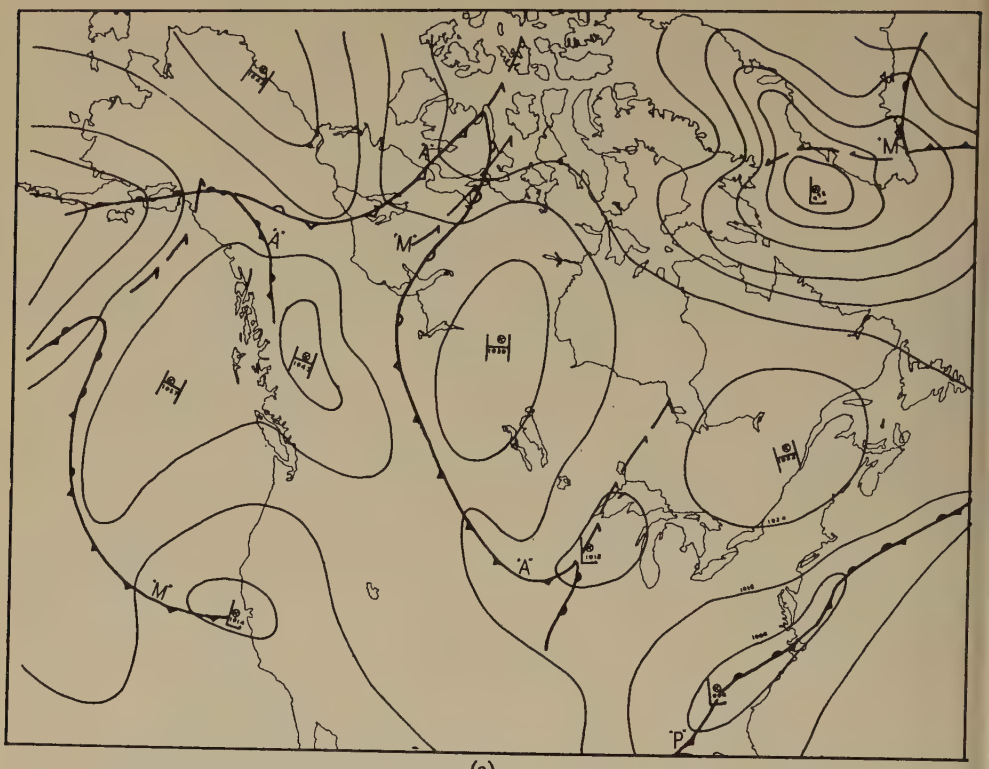
(*a*) *Charts for 0630 and 0300* GCT *11 Jan. 1954 (Fig. 1)*

Over western North America the frontal structure is complex. The polar front is present at upper levels off the west coast. The surface maritime front lies near *Weather Ship P* and this position gives no indication of the important higher-level configuration over north-western Canada. The frontal contour chart shows a dome of mA air over the Canadian Rockies with a pocket of mP air extending below 700 mb over Great Slave Lake. The surface arctic front, for the most part, lies just to the east of the Rockies. There is a weak perturbation on this front to the west of the Great Lakes and a large amplitude one over the North-West Territories accompanying the pocket of mP air. A strong arctic high lies between the two frontal waves. In association with the complex frontal picture the flow aloft shows a split-stream structure and two well-defined jet streams are present.

In the eastern section the three fronts come into proximity with a single westerly stream at 500 mb and a strong jet maximum over Prince Edward Island.

(*b*) *Charts for 1830 and 1500* GCT *12 Jan. 1954 (Fig. 2)*

Marked changes occurred in all areas during the 36-hr period. The dome of mA air moved southward towards the California cold low with deepening of the upper low and intensification of the jet maximum over Oregon where the polar and maritime fronts are fairly steep. The arctic trough moved south-eastward into Hudson Bay with pressure falls of about 24 mb in that area. The mP air was occluded out of that system and the associated jets became quite weak. The cA dome moved from Hudson Bay to the Great Lakes with the surface arctic high showing a similar motion. Ahead of the dome the weak arctic frontal wave has been absorbed in the strong east-coast development. The Georgia wave moved to Nova Scotia and deepened 28 mb with the frontal contour chart showing large amplitude perturbations on all three fronts and the 300-mb chart considerable diffluence over the low.



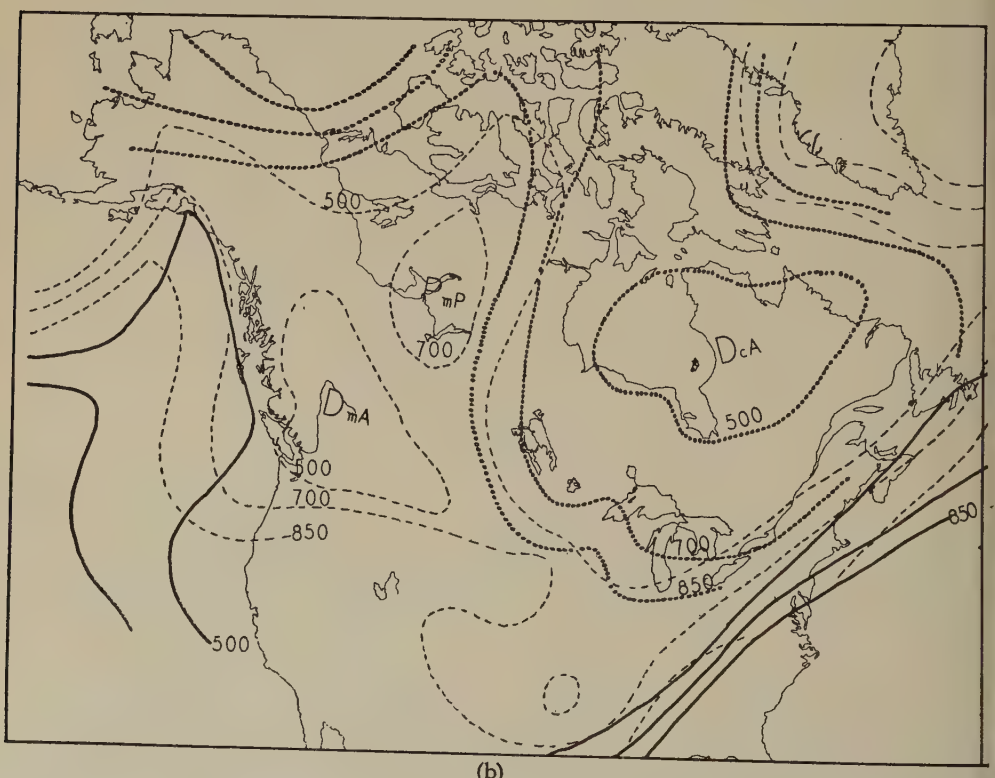
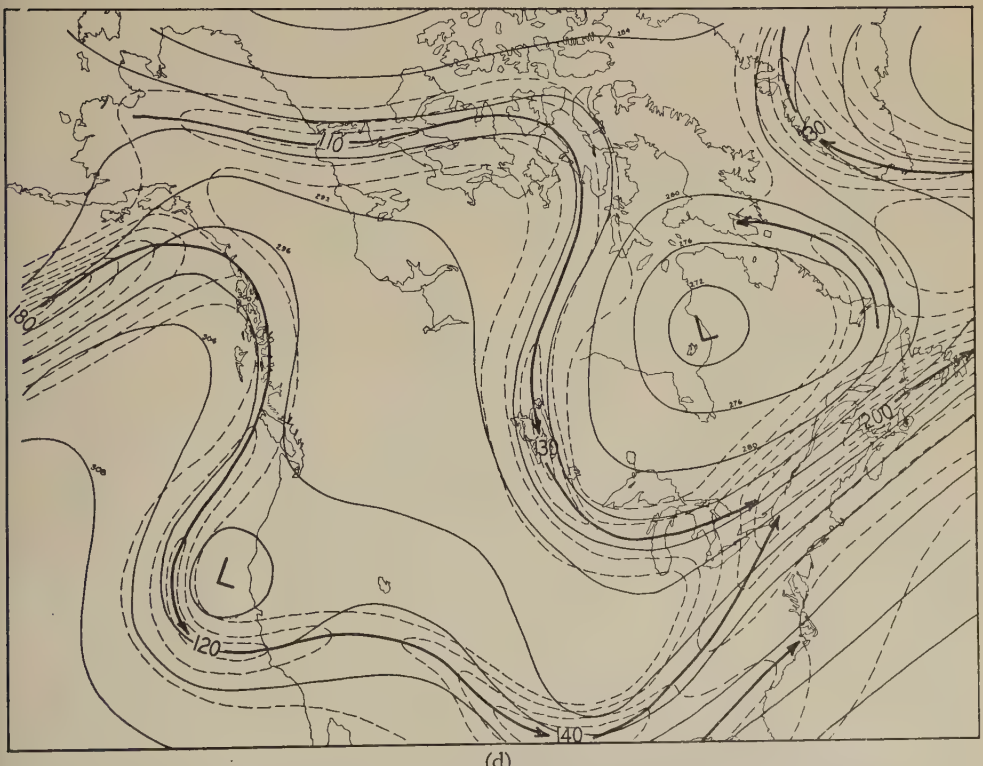


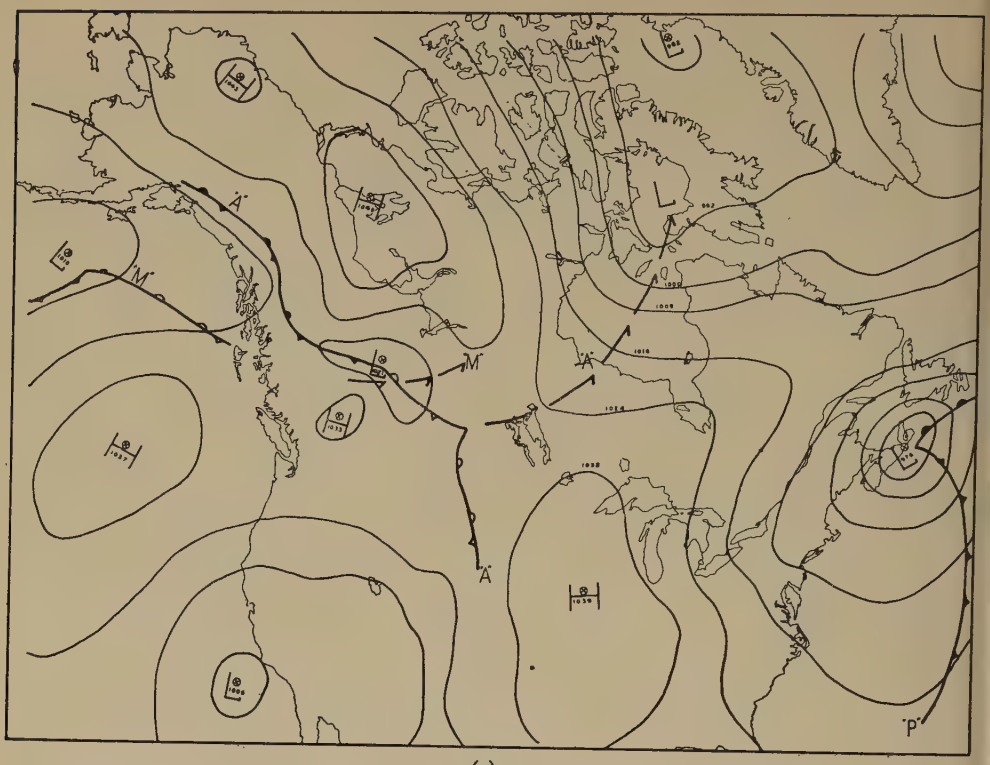
Figure 1. Charts for 0630 and 0300 GCT, 11 Jan. 1954 : (a) surface, (b) frontal contour chart, solid lines – polar front, dashed lines – maritime front, and circles – arctic front.

(c)

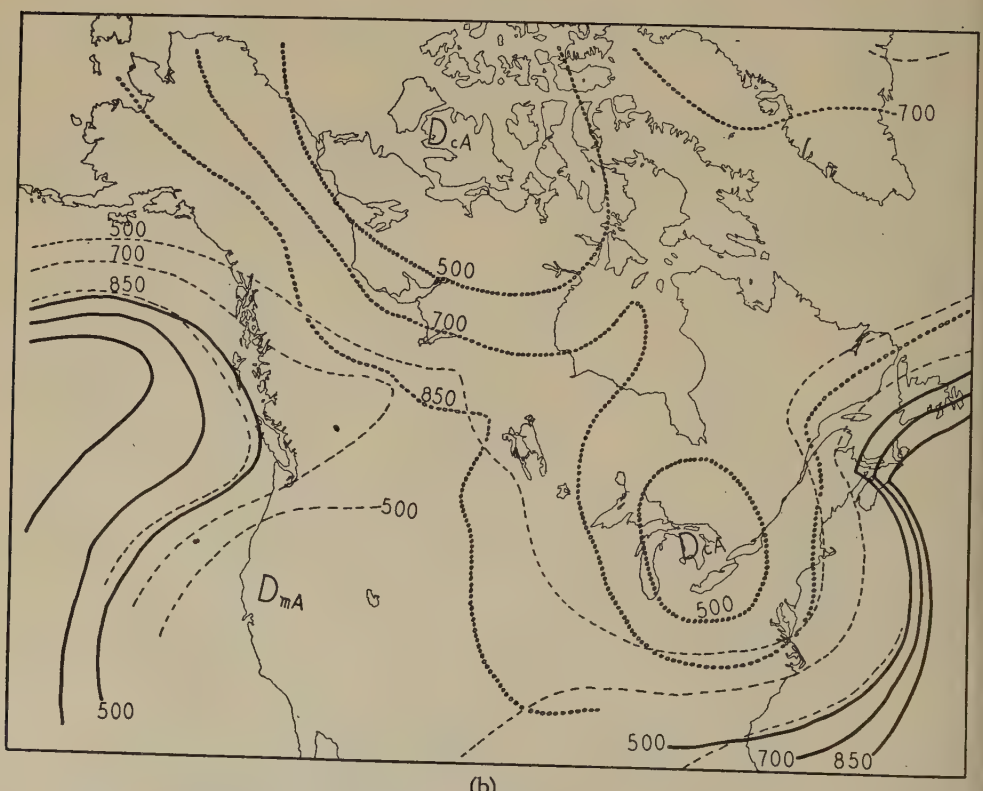


(d)

Figure 1. Charts for 0630 and 0300 GCT, 11 Jan. 1954 : (c) 500 mb, and (d) 300 mb, heavy solid lines – jet axis, dashed lines – isotachs at 20-kt intervals commencing at 60 kt.

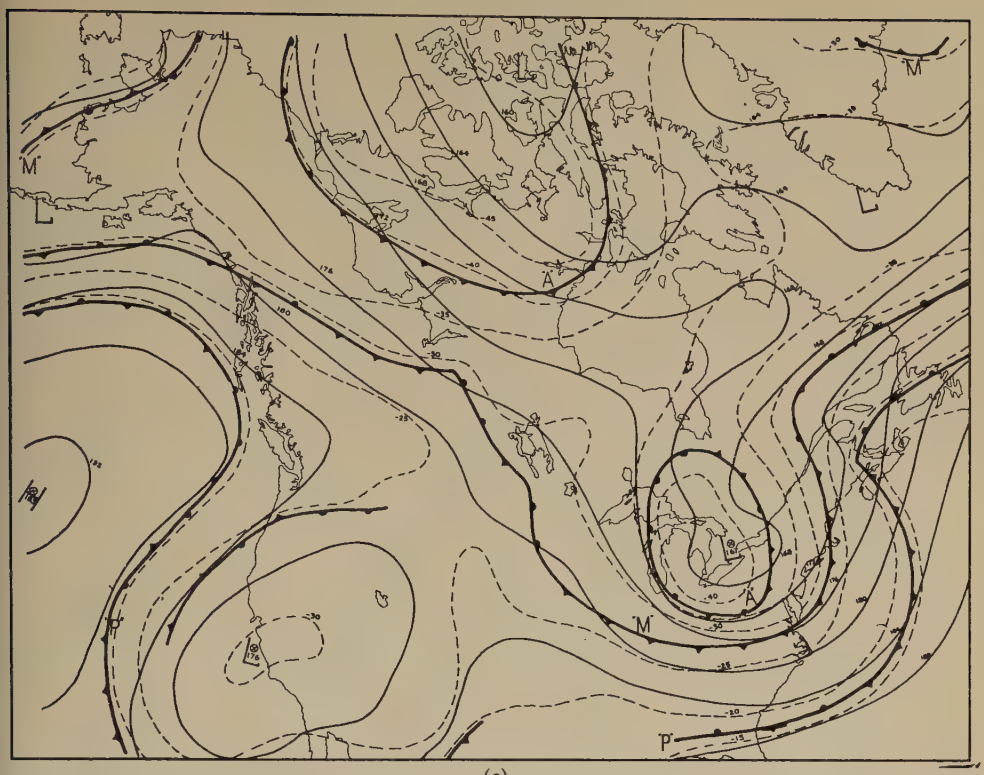


(a)



(b)

Figure 2 Charts for 1830 and 1500 GCT, 12 Jan. 1954 : (a) and (b), same key as for Fig. 1.



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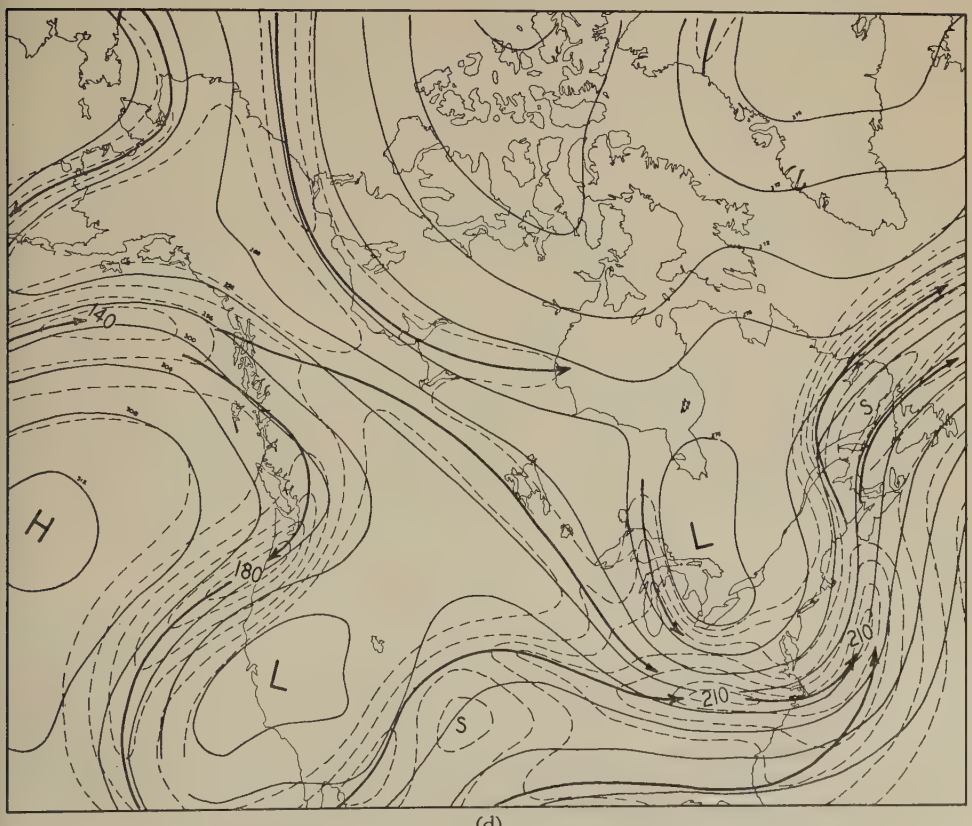

(d)

Figure 2. Charts for 1830 and 1500 GMT, 12 Jan. 1954 : (c) and (d), same key as for Fig. 1.

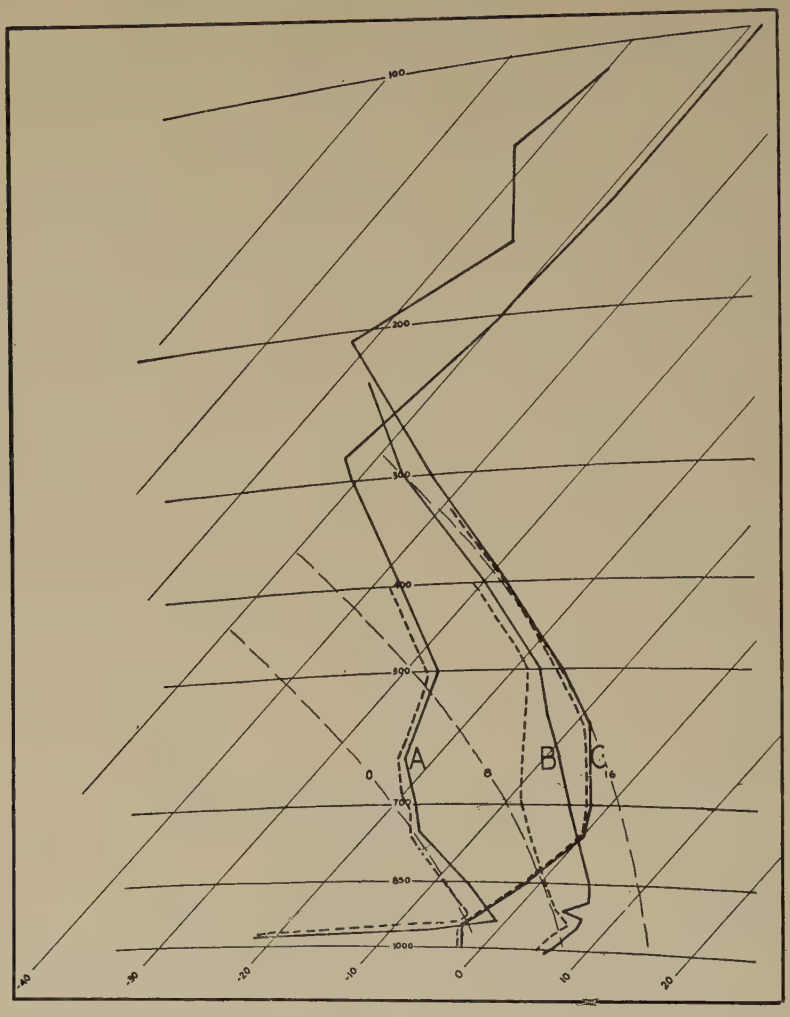

Figure 3. Tephigrams for 0300 GCT, 11 Jan. 1954 : (a) Fort Nelson, B.C., (b) *Weather Ship P*, (c) Washington, D.C. (dashed lines – wet-bulb temperatures).

(*c*) *Tephigrams for 0300* GCT, *11 Jan. 1954* (*Fig. 3*)

The curve for Fort Nelson (945) shows the maritime front at 500 mb with the mA air below overlying a shallow layer of cA air. *Weather Ship P* has the polar front at 500 mb with subsiding mP underneath and mA near the surface. Washington (405) has mT air in the upper levels then a deep hyperbaroclinic zone extending down to transitional cA at the surface; a single frontal surface is taken at 750 mb with a wind increase of 40 kt through the 2,000-ft interval below that point.

(*d*) *Tephigrams for 1500* GCT, *12 Jan. 1954* (*Fig. 4*)

Sault Ste Marie (734) lies within the cA dome. Nantucket (506) shows the maritime front at 590 mb and the arctic front at 770 mb; the lower frontal zone has a wind increase of 30 kt and the upper frontal zone 40 kt. Sable Island (600) in the peak of the frontal wave shows moist mT air over a shallow marine stratum.

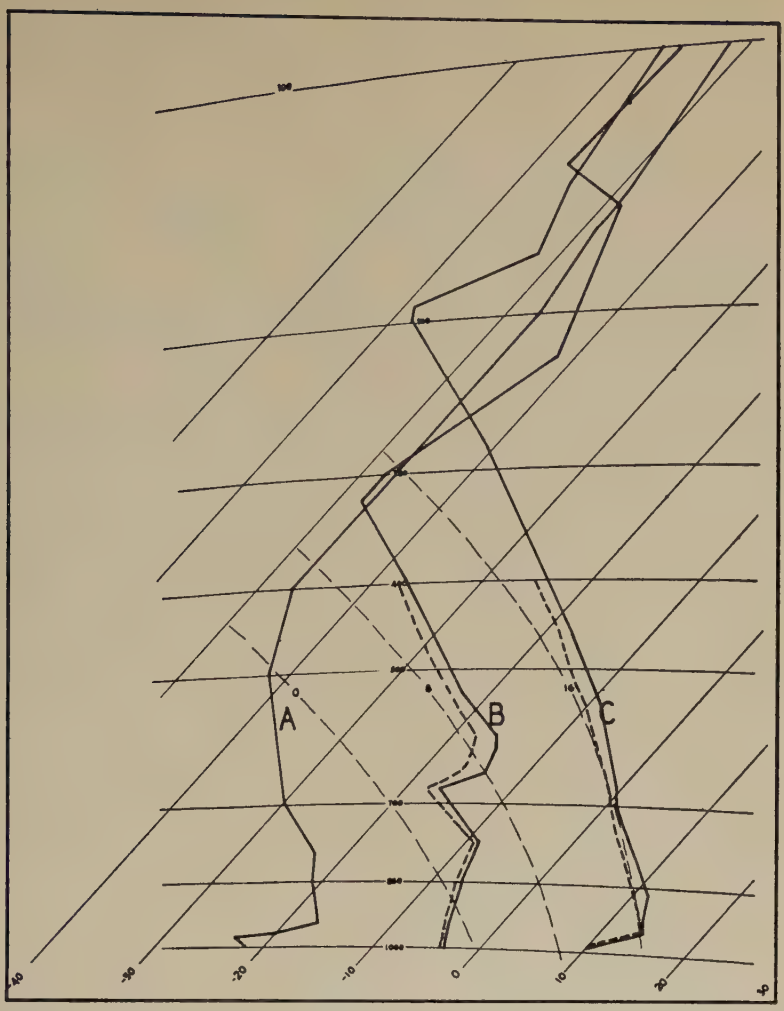

Figure 4. Tephigrams for 1500 GCT, 12 Jan. 1954 : (a) Sault Ste Marie, Mich., (b) Nantucket R.I., (c) Sable Island.

Acknowledgments

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Fifty years of meteorology – some recollections

By R. G. K. LEMPFERT

Formerly Assistant Director, Meteorological Office; past President, Royal Meteorological Society,

In July of this year the Meteorological Office celebrated its centenary and the Royal Meteorological Society passed the hundredth year of its existence in 1950. My personal contact with both institutions has extended over more than fifty years, for I joined the Office in 1902 and became a Fellow of the Society in December 1904. Some of my recollections, especially of the earlier years, may perhaps be of interest to the present Fellows and it is with pleasure that I respond to the Editor's invitation to contribute an article to the Quarterly Journal for my eightieth birthday.

In 1902 the Office was housed at 63 Victoria St., in what is now probably the oldest building still standing in that thoroughfare, and the Society occupied a rather more modern residential flat at No. 70. Both institutions were already rather cramped for space to house their growing libraries. The Society had no adequate meeting room and our monthly meetings were held at the Institution of Civil Engineers in Gt. George St. At an earlier date our home had been in Gt. George St., but the building had been demolished to make way for the Institution's new premises; hence the connection.

Two world wars and the growing demands of flying have left their mark on both Office and Society. I have seen Fellowship of the latter increase from 658 in 1904 to 2022 in 1954. The Office numbered about 40 (including the messengers), and now the figure stands at more than 3,000, having shrunk to that from a maximum of about 6,000 during the late war.

The changes have not been merely in numbers. Formerly the Fellowship of the Society was preponderatingly amateur for the very good reason that there were very few professional meteorologists, i.e. people who earned their living in meteorology. I think you could have counted their number, apart from the Office staff, on the fingers of your hands : one hand might even have been enough. There were of course many professional scientists sufficiently interested to take up Fellowship but the amateurs *pur sang* were greatly in the majority, and we owe a lot to them.

Many made regular observations which they reported to the Society, which was thus responsible for a network of observing stations for which it published results in the *Meteorological Record*. The stations were inspected at intervals by the Assistant Secretary, Mr. William Marriott, who was responsible for the scrutiny of the observations and their summaries. A selection of these was passed to the Office for publication in the annual volumes of *Returns from Stations of the Second Order*. That arrangement lasted until 1911, when the Office took over and incorporated the summaries in its much enlarged *Monthly Weather Report*, which Fellows are still entitled to, if they wish to receive it. Many of the observers had equipped their stations with barographs and the Office was able to borrow the records and use them in the preparation of the hourly maps for selected occasions discussed in the *Life History of Surface Air Currents*.

The preponderatingly amateur nature of the Fellowship was reflected in the discussions and the contents of the *Journal*, which were mostly of a climatological or statistical character. A whole evening was devoted every year to discussing the phenological report and generally attracted a better house than usual. I can remember meeting Fellows from the country regularly on these occasions who never attended otherwise. Simpson and I, fresh from the atmosphere of a university research laboratory, sometimes chafed at the absence of ' real science ' from our proceedings, but we need not have have distressed ourselves. In the course of time, physical and mathematical conceptions and even advanced mathematics have found their way into the *Journal*, in a measure to satisfy the most exacting of high-brows. On looking through the back numbers of fifty years ago I am struck by

ıe number of papers of lasting importance they contain. For instance there is the work f W. H. Dines and the Wind-force committee which gave us the pressure-tube anemo-ıeter and all that that has meant for anemometry.

That was just before my time which, however, saw the early stages of upper-air work ı this country. One of my earliest meteorological recollections is of a visit to W. H. Dines t Oxshott, where I found him in the middle of a kite ascent on an overcast drizzly day.)uring the last decade of the previous century the investigation of the upper air with ites and balloons had got under way on the continent, but this country had lagged ehind. Glaisher and Welsh's pioneer work of the 50's had not been followed up. Attempts to do so had received a rude check when a member of Parliament who had olunteered to participate in a meteorological balloon flight was carried out to sea and ever heard of again. With the turn of the century things changed, but Shaw never ucceeded in persuading the Government or private interests to put up money for upper-air ork on anything like the scale which seems to have been accepted as a matter of course ı the Kaiser's Germany. Nevertheless the combined efforts of Office, Society, British Association and private enthusiasts like Dines and Cave made a very substantial con-ribution in the years that preceded the outbreak of the first world war. The work one at Manchester University (under Schuster's direction) at its kite station on Glossop Moor must also be mentioned. Much of this work was published in the *Journal* and here was no longer much basis for our fault-finding.

At one of the first Council meetings I attended, the question of an effort to increase ellowship and general usefulness of the Society arose. Rather diffidently I put forward he suggestion of holding occasional meetings away from London for the benefit of country ellows. The idea caught on and in due course a meeting was arranged in Southport, here the Corporation maintained a station of the first order, equipped with all the ssential recording instruments, in the charge of the Borough meteorologist, Mr. J. axendell. It was not the first time meteorology had visited Southport, for a meeting f the International Meteorological Committee had been held there in 1903, simultaneously ith a meeting of the British Association. Baxendell and the Corporation made us very elcome and the meeting was a great success. The Corporation supplied transport ı the shape of a char-a-banc drawn by two stalwart horses to take us to the anemometer tation at Marshside, where one of the first pressure-tube anemometers used in this ountry had been installed. Its exposure amid the sandbanks and marshes of the Ribble stuary was well nigh ideal : no near obstacles of any importance but the mountains of akeland could be seen on a clear day.

The experiment of a provincial meeting was repeated at Manchester University ı 1910, where, as I have already mentioned, Schuster had kindled the meteorological orch, and was again a success. The Manchester physics staff contributed papers on tmospheric electricity and on work arising out of their registering-balloon ascents and orless and I read a paper on line squalls (' *live* squalls,' according to the local reporter) hich set out the results of investigations carried out in the Office. Meetings away from ondon are now part of our routine activities.

In 1910 the Office left Victoria St., and took up its quarters in the specially designed uilding at South Kensington, destined so soon to become too small for all its new activities. he following four years brought many new interests, the Iceland cable, the first attempts) use wireless to collect information from ships at sea, G. I. Taylor's kite-flying activities ff the Banks of Newfoundland which had such unexpected repercussions on the question f atmospheric turbulence, etc., and then came the 1914/1918 war.

Its first effect on the Office was to bring the work of the forecast division, of which was Superintendent, to a standstill. The Post Office wires on which we depended for

our reports were overloaded with other Government business and that was that. C the evening of the second day a uniformed Admiralty messenger appeared at my priva address and insisted on my accompanying him to the Admiralty. There it was explain to me rather brusquely that people at the Admiralty had something more important to c than to worry about weather reports. I never found out what had given rise to th summons. I presume that the Hydrographer, who was our link with the Admiralty, ha realized our predicament and had asked somebody to look into the matter.

The situation improved gradually and, after an interval, the reports from the hom stations and a few stragglers from France and the neutral countries round the North Se came in fairly regularly. Presently it was realized that weather reports from this countr might be of use to the enemy and we received instructions to cease all publication an treat everything as confidential. Some time however went by before any steps were take to make meteorology of more use to our own side. Then one day there came a deman from the Admiralty for daily forecasts for the eastern Mediterranean. Shaw met tha emergency by asking the War Office to make Major Henry Lyons, who had been Directo of the Egyptian Survey Department which included the Meteorological branch, availab for special duty with us. Lyons was of course a regular officer, R.E., and his recall to activ service when war broke out had been automatic, but up to that time the War Office ha found no more suitable work for a geographer-meteorologist of his eminence than to assig him to Chatham in charge of recruiting. The request was granted immediately. Knowin all the ropes, Lyons got us reports from a few Mediterranean stations and we proceede under his guidance to issue the required forecasts. (I mention these details because they wer so typical of the attitude towards meteorology which prevailed in high quarters at that time How completely that attitude had changed by the end of the war is common knowledge).

The war was a difficult time for the Society as for other bodies. Many Fellows wer called up or undertook war work and their services were temporarily lost to the Society The President, Mr. Cave, held a commission in the Meteorological Section R.E. and th Assistant Secretary, Mr. Hampton-Brown, who had succeeded Mr. Marriott in 191 joined the Naval Meteorological Service. Lyons, now in temporary charge of the Office succeeded Cave in the Presidental chair in 1915 and it was largely due to his drive that w came through as well as we did. We even managed to produce in 1917 a volume of th *Journal* of more ample proportions than ever before. At the close of the war a seriou housing difficulty had to be faced. A rather staggering increase in the rent demande for 70 Victoria St., and the lack of space for the growing library rendered a move almos imperative. A plan worked out jointly with the Cambridge Scientific Instrument Compan under which the Society, or rather a private company ' Meteorological Estates,' was t acquire a house in Grosvenor Gdns., and accept the Society and the Instrument Compan as tenants, fell through at the last moment. Soon after we acquired our present home a 49 Cromwell Road. The capital for the purchase was subscribed by some of the Fellow on terms that were very generous to the Society. For several years the Hon. Treasurer Mr. F. Druce, made it his business so to nurse the finances that the loan could be paid off. In that laudable desire he succeeded before he resigned the Treasurership, thank again to the generosity of some of the lenders.

Developments during the inter-war years will be known to many present Fellows Many of them have had an active part in them. My personal recollections can be o little interest. By the time the second world war started I had reached the retiring age What of the future ? How far will our knowledge of the atmosphere advance ? Ca we hope for anything approaching astronomical accuracy in our forecasts, short-term o seasonal ? Further outlook – none issued, but the prospects for further useful wor seem bright both for Society and Office.

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Spectra and cross-spectra of velocity components in the mesometeorological range

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(Manuscript received 8 July 1955)

SUMMARY

Power spectra of horizontal wind components at 91 m indicate the existence of a gap near wavelengths of the order of 10 km or periods of the order of one hour.

1. General properties of spectra in the micrometeorological range

Considerable information is becoming available regarding spectra and cross-spectra of velocity components in the micrometeorological range, covering periods from a fraction of a second to several minutes, and heights above the ground from a few tens of centimetres to 100 m (see Robinson 1953, Sheppard 1952, MacCready 1953, Cramer and Record 1955 and Panofsky and McCormick 1954). The general conclusions drawn from these studies indicate that :

(a) Local isotropy exists and the spectral law postulated for the inertial subrange is satisfied at high frequencies.

(b) Toward lower frequencies, the spectra of the horizontal velocity components exceed those of the vertical velocity.

(c) Mechanical turbulence is more important at higher frequencies than convection.

(d) The scale of turbulence increases with height.

(e) As expected from (a), the smallest eddies contribute little to the stress.

2. Distribution of kinetic energy and stress over the various frequency ranges

Studies of the distribution of kinetic energy over eddies with periods of 12 hr or greater, or wavelengths exceeding 200 km, are now being begun at various places, but have not yet been published to the authors' knowledge. Such estimates can of course be made from synoptic observations. Between the micrometeorological and synoptic ranges lies the mesometeorological range covering periods too short to be studied from synoptic observations and too long for most micrometeorological series. This range has considerable importance in the estimate of stability of meteorological observations, and for a statistical treatment of diffusion.

Several general statements can be made in regard to the relative importance of the contributions of micrometeorological variations to the total kinetic energy :

(a) In the case of vertical motion, the micrometeorological range contributes by far the largest fraction of the total energy.

(b) For horizontal velocity components, the energy in the micrometeorological range is only a small fraction of the total.

These two points are illustrated in Table 1 which compares the variance of five-second deviations from hourly averages with the variance of hourly averages. All observations were made on 91 m at the Brookhaven 125-m tower, the instrumentation of which has been described previously (e.g., Panofsky and McCormick 1954).

* Now with U.S. Weather Bureau.

TABLE 1. VARIANCES AND COVARIANCES IN BROAD SPECTRAL REGIONS

	Five-second average deviations from hourly means (m^2 sec^{-2})	Hourly means (m^2 sec^{-2})
$\overline{w'^2}$	1·2	0·03
$\overline{u'^2 + v'^2}$	6·5	35·7
$\overline{u'w'}$	− 0·7	+ 0·1
$\overline{u'v'}$	+ 0·4	+ 1·3

Here, w is the vertical velocity, u and v are horizontal wind components. In the first column these components refer to the direction of the mean wind, and to the direction at right angles to the mean wind. In the second column u is measured in the W-E direction, v in the N-S direction. A bar means a time average, a prime a deviation from such an average.

The data in Table 1 were observed at 91 m, during different periods. For example, all the information in the first column is derived from a period only about one hour in length. However, the magnitudes are typical for many other periods as well (Panofsky and McCormick 1954). The data in the second column were derived from a period a week long.

In addition to the points made above, the table shows that the small eddies contribute little to the meridional flux of zonal momentum, confirming an assumption made by many investigators of the angular momentum budget (e.g., Widger 1949). Finally, the table indicates that the small eddies produce most of the vertical flux of horizontal momentum at least at 91 m. At higher levels, on the other hand, the vertical flux of zonal momentum is often ascribed to eddies of cyclone size (see White and Cooley 1952).

3. POWER SPECTRA IN THE MICROMETEOROLOGICAL RANGE OF FREQUENCIES

Only the energies of the horizontal velocity components possess important contributions both in the synoptic and the micrometeorological frequency range. It is of some interest to determine whether the energy spectrum possesses a gap between these ranges. Fig. 1 summarizes a large number of spectral estimates in the micrometeorological range.

The abscissa is the logarithm of the wave number, k, assuming that $k = n/\overline{V}$ where n is the frequency, and $\overline{V}$ the mean wind speed. In order that the area in Fig. 1 represent contributions to the variance of horizontal velocity components, the spectral estimates plotted were first multiplied by the frequency. The resultant variances were then divided by the square of the mean wind speed; this division produces a non-dimensional ordinate which appears to possess similar characteristics for different wind speeds, and depends mainly on the stratification.

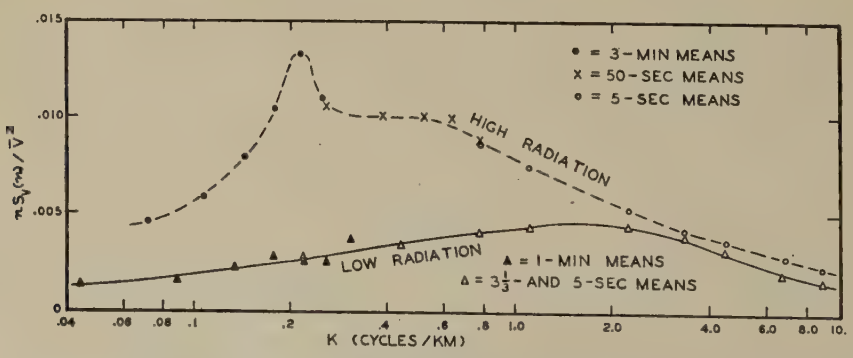

Figure 1. Behaviour of the spectra of horizontal wind components under high and low radiation conditions in the micrometeorological frequency range.

The two curves in Fig. 1 illustrate the behaviour of the spectra of horizontal wind components nder clear and cloudy skies, respectively. Observations in the high-frequency portion of each urve were derived from the composite spectra published by Panofsky and McCormick (1954). pectra of wind speeds and wind components in the mean wind direction were used interchange- bly, since no noticeable difference in their properties was shown to exist in the micrometeoro- ogical range. In each case, the original observations consisted of eye-averages of the wind omponents over 3½ to 10 seconds. Such averages could be estimated since the Esterline-Angus race moved at the speed of 0·05 in./sec. The averaging process was assumed to produce rithmetic means, so that the spectrum of the wind components could be estimated from the pectrum of the averages by :

$$S_V(n) = S_{\bar{V}}(n)\left(\frac{n\pi T}{\sin n\pi T}\right)^2 \quad . \quad . \quad . \quad . \quad (1)$$

vhere T is the averaging period. The low-frequency portions of the spectra were derived from eriods in which observations were available over 9 and 2 hr, respectively. In each case, the riginal observations were averaged over 50 sec or more, as indicated in the figure. Again, Eq. (1) was used to correct for the filtering produced by the averages. The technique for putting ogether portions of spectral estimates covering different frequency ranges was discussed by Griffith, Panofsky and Van der Hoven (1955).

Fig. 1 shows the previously-noted tendency of heat convection to produce energy at lower frequencies than mechanical turbulence. Of particular interest to this discussion, however, is the tendency for both sets of spectral estimates to decrease with decreasing value of k at the ow-frequency end of the figure. Particularly, little energy at 91 m is generated near $k = 0{\cdot}1$, or at a wavelength of 10 km.

4. Spectra in the mesometeorological range

Fig. 2 shows power spectra at 91 m for various horizontal velocity components, derived from observations covering 39 hr and a week, respectively. The ordinate and abscissa are the same as those used in Fig. 1. In contrast to the result shown in Fig. 1, those in Fig. 2 were based on Esterline-Angus traces moving at the normal rate of 0·05 in./min. Thus, averages over 5 min could be estimated by eye. Spectra of the 5-min and 20-min averages were then treated by Eq. (1) to yield the spectra in Fig. 2.

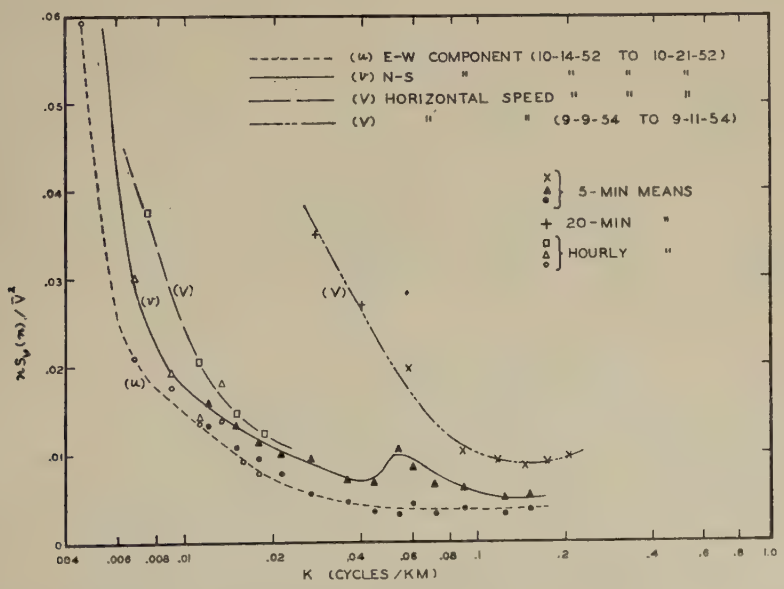

Figure 2. Behaviour of the spectra of horizontal wind components in the mesometeorological frequency range.

Again, low spectral estimates appear near $k = 0{\cdot}1$, increasing sharply from there towar lower frequencies. On combination of the results from Figs. 1 and 2, it is seen that spectr of horizontal wind components are characterized by a relatively low magnitude near $k = 0{\cdot}1$, c near periods of the order of an hour. The conclusion that a ' gap ' exists in spectra of horizonta velocity components must, however, be considered to be highly tentative. It is based on relativel few observations and on an interpolation between different sets of data; more conclusive result would be obtained from a long run of one-minute average observations.

If the minimum in the spectrum near 1 hr or 10 km is real, it must be due to the absenc there of physical processes : convection and mechanical turbulence produce energy at shorte periods (at least at 100 m and below), and motions of cyclonic scale as well as diurnal variation have, of course, much longer periods.

The spectral gap suggests a rather convenient separation of mean and turbulent flow in the atmosphere : flow averaged over periods of about an hour (or smoothed to correspond t such an average) is to be regarded as ' mean ' motion, deviations from such a mean as ' turbu lence.' With this convention, heat convection would be included in the group of turbulen motions. This division is useful for diffusion problems; for hour-averages can be estimated b smoothing synoptic information which can be analyzed in detail. ' Turbulence ' superimpose on this motion must be treated statistically. This separation implies the assumption that a ga in the Lagrangian spectrum exists in the same frequency region as the gap in the Eulerian tim spectrum, an assumption not yet tested.

Finally, the spectral gap near one hour suggests that wind observations should be average over at least several tens of minutes before stable estimates can result. This explains why th two-minute averages commonly used in the reduction of pilot-balloon runs are characterized b pronounced ' turbulent ' irregularities.

Acknowledgments

The authors wish to thank the meteorological personnel at the Brookhaven National Laboratory for making available the records on which this analysis is based, and to Mr. James Cooley of the Institute for Advanced Study who supervised many of the computations.

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551.508.71

A method of increasing the range of the Dobson-Brewer frost-point hygrometer in jet aircraft

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(Manuscript received 2 June 1955)

1. Introduction

The measurement of humidity in the troposphere and lower stratosphere has been carried out for a number of years by the Meteorological Research Flight using the Dobson-Brewer frost-point hygrometer (Brewer, Cwilong and Dobson 1948). There is little difficulty in recognizing the frost deposit as the air passes over the cooled thimble of this instrument at temperatures above — 50°F (— 45°C) but as the temperature is lowered the thickness of the deposit decreases until it requires an experienced operator to obtain readings below about — 100°F (— 73°C). At temperatures below about — 120°F (— 85°C) any frost deposit may be glassy in form and difficult to observe without special optical arrangements.

Over the United Kingdom the frost point is usually about — 100°F (— 73°C) at 40,000 ft (Bannon, Frith and Shellard 1952) and — 120°F (— 85°C) at 50,000 ft (Murgatroyd, Goldsmith and Hollings 1955). There is strong evidence that similar values will be found at comparable heights over the tropics and polar regions.

It was suggested by A. W. Brewer that the effective range of this instrument could be increased if it were supplied with compressed air. A compression to surface pressure from an altitude of 50,000 ft would increase the measured frost point by over 20°F (11°C), an amount which may be sufficient at present-day aircraft ceilings. This can be accomplished in a jet aircraft by taking air from the engine compressor where the compression ratios of 6 or 8 may be available.

2. Installation and operation

The installation is shown in Fig. 1. Air from the engine compressor is fed through a control valve to the hygrometer which is located in the pressurized cabin. Inlet and outlet pressures are measured and valves provided to select either the outside air or air from the compressor, as desired. If the control valve is adjusted to obtain surface pressure within the hygrometer the value of frost point obtained is conveniently corrected to its equivalent value at the aircraft's altitude by means of a tephigram or other equivalent thermodynamic diagram. Several flights were made comparing the frost point obtained directly with the outside air with that obtained using the compressed air at different pressures and it was verified that the same result was obtained.

One of the advantages of this system is that the air within the hygrometer is at a higher pressure than the air in the pressurized cabin. Hence any leaks will be outwards. If outside air

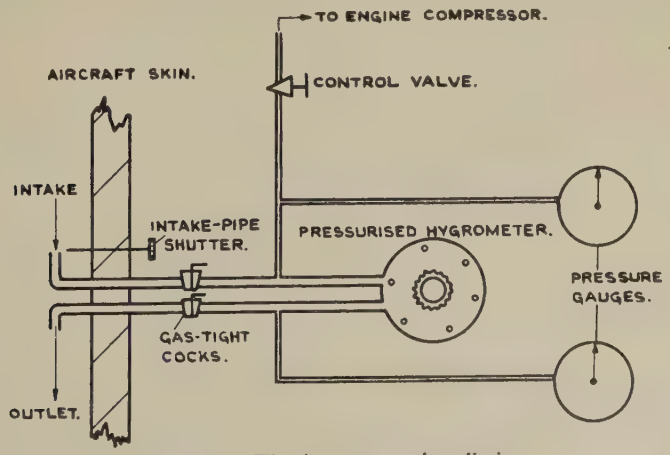

Figure 1. The hygrometer installation.

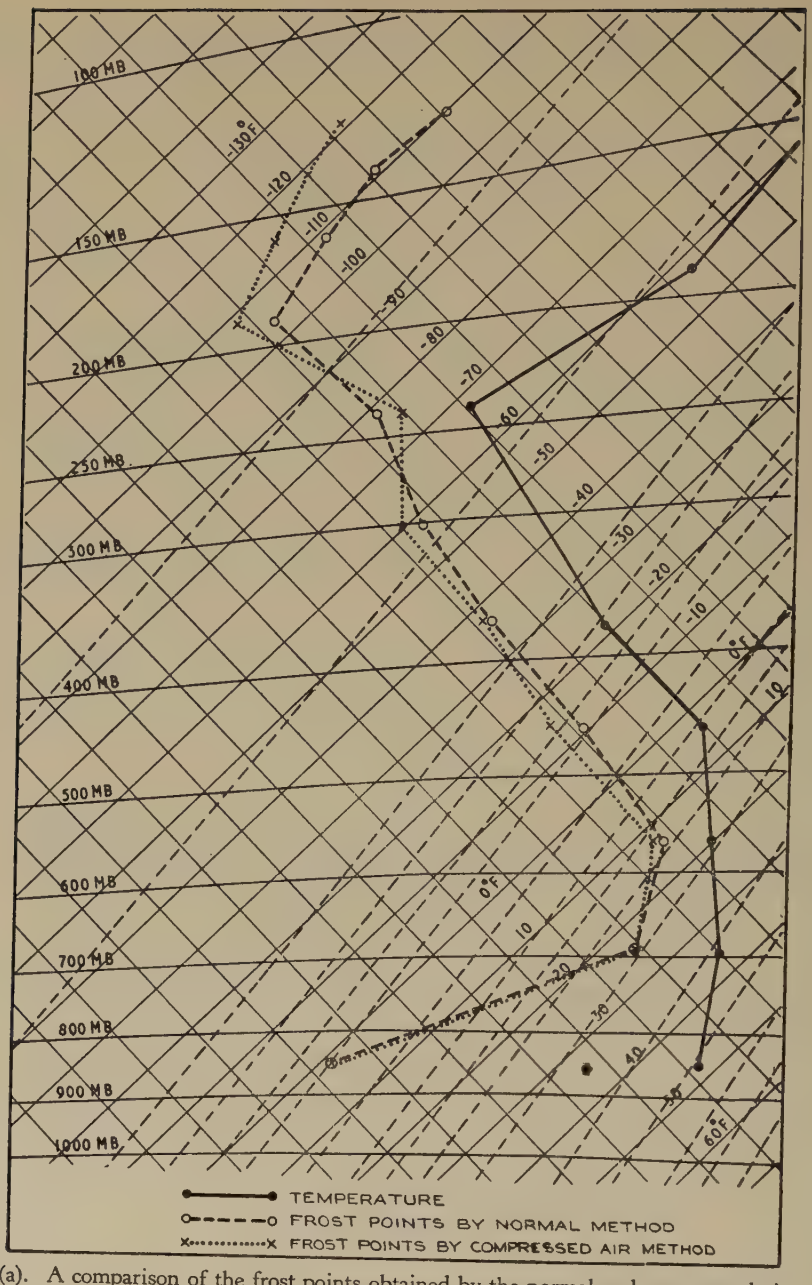

Figure 2 (a). A comparison of the frost points obtained by the normal and compressed-air methods. With a probable small leak.

is used the pressure is lower within the hygrometer and leaks will be inwards. At very low humidities a small leak of this nature causes a large error in the measured frost point. Figs. 2 (a) and 2 (b) show a series of temperature and humidity measurements to 50,000 ft plotted on a tephigram comparing the two methods of measurement. In the case plotted in Fig. 2 (a), the hygrometer had a slight leak which resulted in the divergence of results at the highest levels.

The compression of air produces also a considerable increase in its temperature with the result that a hygrometer used in this way requires an efficient cooling system for its thimble and good heat insulation. The system should not be used at low levels for this reason and the change-over from outside air to compressed air is usually made by the Meteorological Research Flight at an altitude of about 30,000 ft.

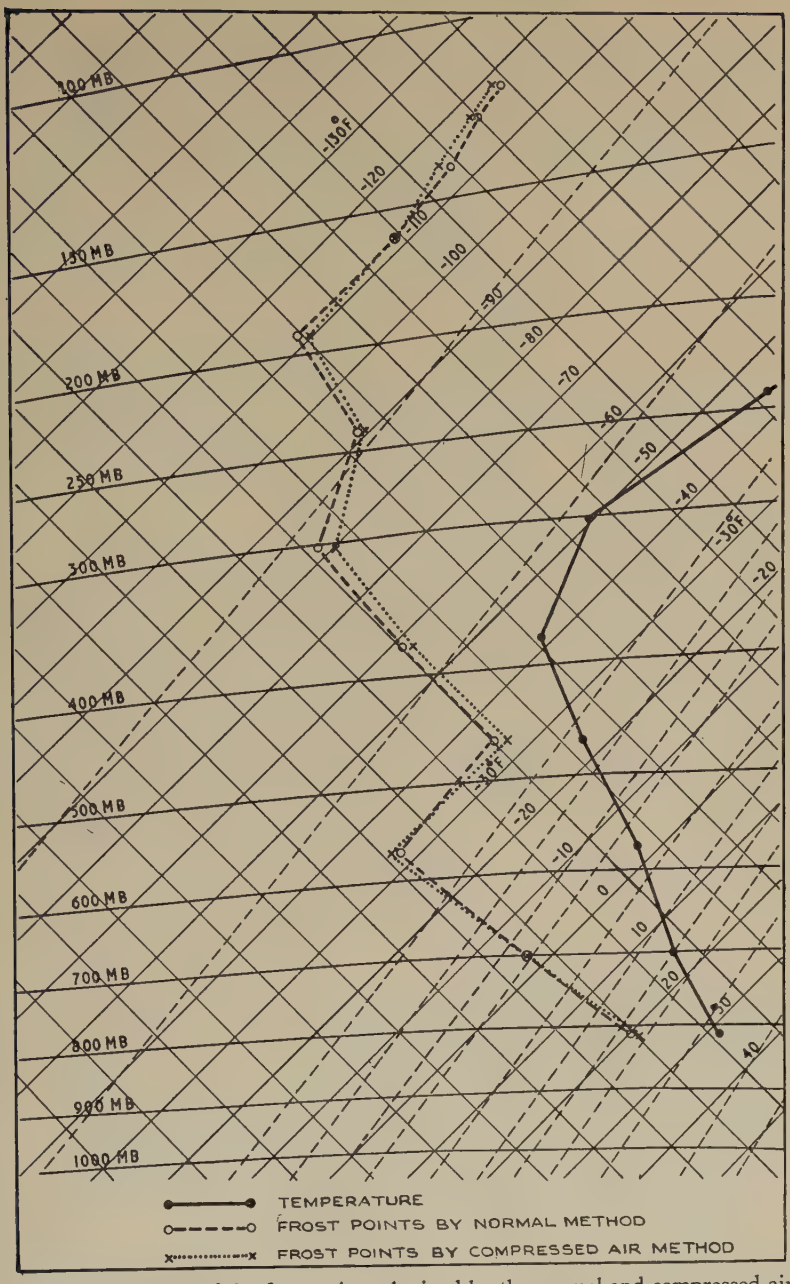

Figure 2 (b). A comparison of the frost points obtained by the normal and compressed-air methods. With all seals satisfactory.

Acknowledgment

The author is indebted to the Director, Meteorological Office, for permission to publish this note.

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551.577.11

The break-up of large water drops falling at terminal velocity in free air

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(Manuscript received 25 May 1955)

1. Introduction

Heavy rain frequently forms in tropical cumulus clouds which lie wholly below the freezing level. In maritime air containing high concentrations of large hygroscopic nuclei, it may be possible to account for the observed intensities of rainfall on the basis of one nucleus producing one raindrop. In the interior of continents it seems necessary to postulate a process of drop multiplication by the break-up of drops which have grown by coalescence beyond the limit of stability.

Langmuir (1948) examined the conditions of cloud water content, updraught velocity, and vertical thickness under which growth of droplets to instability could take place. However, the only data then available on the number and size distribution of the fragments produced by the break-up of unstable falling drops were those of Lenard (1904). This stimulated Blanchard (1950) to undertake further studies of the mechanism of droplet break-up. Although he did not pay special attention to the size distribution of the fragments, he noticed that the degree of turbulence of the air had a marked effect on the size at which break-up occurred.

Both Lenard and Blanchard worked with vertical wind tunnels. These have the advantage that the droplets may be kept under continuous observation while falling relative to the air, but suffer from the fact that the turbulence regime may be very different from that obtaining in the atmosphere. We therefore thought it worth while to study the break-up of water drops falling in free air.

2. Technique

Large drops of distilled water were allowed to fall from various heights up to 20 m from the tower of the Command and Staff College, Quetta. The fragments of those which broke up were caught on sheets of filter paper.

Much difficulty was experienced in producing regular single drops of large size. After several attempts with more complicated apparatus, we used a simple glass tube of internal diameter 6 mm fitted with a rubber bulb at one end. According to the size of droplet required, a quantity of water was drawn into the tube and was then ejected by a sudden pressure on the rubber bulb. Occasionally the large drop formed in this way was accompanied by one or more smaller ones, but in these cases no attempt was made to capture the fragments.

It was not easy to produce drops of uniform size by this method, and so the volume of each parent drop was calculated subsequently by adding together the estimated volumes of the fragments.

The height from which the drops were allowed to fall was chosen so that break-up normally occurred a few metres above the ground, when all the fragments could be caught on a filter paper approximately 60 × 100 cm, stretched on a wooden frame. From above, it was easy to see when some fragments missed the filter paper, and in this case the observation was rejected. Wind during the experiments was generally light, less than 5 kt. We did not find it necessary to treat the filter paper with a dye, as had been done by previous workers (Lenard 1904; Marshall, Langille and Palmer 1947; Anderson 1948), who used this technique for the analysis of raindrop sizes, since in this case we could measure the spots immediately after impact. From preliminary studies of the rate of spreading, and of the evaporation of the spots, we adopted the procedure for measurement which was least subject to error on this account. As soon as possible after impact, the observer marked with a pencil the circumference of each spot in order of size from the smallest to the largest, so that the former should not disappear by evaporation, and the latter should have time to grow to full size, before marking. Measurement of the spot diameters could then be made at leisure. In cases where the spots were not circular, we adopted the mean of the diameters in two perpendicular directions.

The conversion of spot diameters to those of spherical droplets producing them was made with the help of a calibration curve appropriate to the filter paper used in the experiments. Drops of known volumes (equivalent spherical diameters ranging from 2·3 to 6·5 mm) were allowed to fall on to the paper from heights sufficient for the droplets to attain their terminal velocity. The latter precaution was necessary since it was found that the size of the spots produced was influenced by the impact velocity of the drops. The smaller drops were produced with hypodermic needles, and the larger ones from burettes fitted with rubber-tube adapters. About 12 drops of each size were used to determine the corresponding mean spot diameters.

Within the range of drop sizes used in calibration, the relation between drop and corresponding spot diameters was found to correspond very closely to :

$$D = 0{\cdot}47\, S^{2/3} \quad . \quad . \quad . \quad . \quad . \quad (1)$$

The index 2/3 was to be expected from geometry, while the figure 0·47 is a constant depending on the type of filter paper used. We therefore felt fairly confident in extrapolating the calibration curve to cover the range of drop diameters from 1 to 10 mm.

3. Results

(*a*) *Critical size for break-up*

Blanchard (1950) found that the size at which droplets became unstable depended on the degree of turbulence of the air through which they were falling. This may explain the fact that in our experiments the smallest drop observed to break up during a fall of 20 m in free air had a diameter of approximately 8 mm. In wind tunnels the critical diameter appears to be in the neighbourhood of 5·5 mm (Lenard 1904, Blanchard 1950). On the other hand, the largest raindrops so far recorded had diameters between 6 and 7 mm.

Break-up may depend upon random encounter with air eddies, and it is possible that if we had been able to work from a greater height we should have observed the break-up of smaller drops. However, this distinction loses significance in clouds since the drop will continue to grow while falling. It is difficult to say whether in the wind-tunnel experiments or in those reported here the conditions of turbulence more closely approached those likely to be encountered in cumulus clouds. Furthermore, we cannot say what may be the effect on the stability of a drop of the numerous collisions with small droplets which it must encounter when falling through cloudy air.

(*b*) *Number of fragments*

The diameters of the parent drops ranged between 8·5 and 12·5 mm, though the great majority, 94 in number, lay between 9·0 and 12·0 mm. The number of fragments produced by individual drops varied from 3 to 97. In Fig. 1 are shown frequency histograms of the number of fragments produced per parent drop. There is some indication in the larger size-groups of two maxima in frequency, suggesting that different break-up mechanisms may be responsible for the production of small and large numbers of fragments. The number of experiments is too small for us to draw any conclusions on this point. It is clear, however, that the number of fragments tends to increase with the size of the parent drop.

(*c*) *Size distribution of fragments*

The size distribution of the fragments is of interest to the cloud physicist who is concerned to know how quickly they may grow to the limit of stability and so break up in their turn. Their size distribution obviously depends on their number, but we have preferred here to group them merely according to the diameter of the parent drop. What is of main interest is the net result of the break-up of a large number of droplets in each size group.

Fig. 2 shows the size-distribution histograms for the three groups of parent-drop diameters. It appears that while the number of fragments increases with increasing parent-drop diameter, the form of the size-distribution curve of the fragments remains unchanged except for a gradual extension towards the larger sizes. There is some slight evidence for a secondary maximum. The larger drops tended to give on break-up a few droplets of 6-8 mm diameter accompanied by about 30 small droplets, mainly of diameter 1-3 mm.

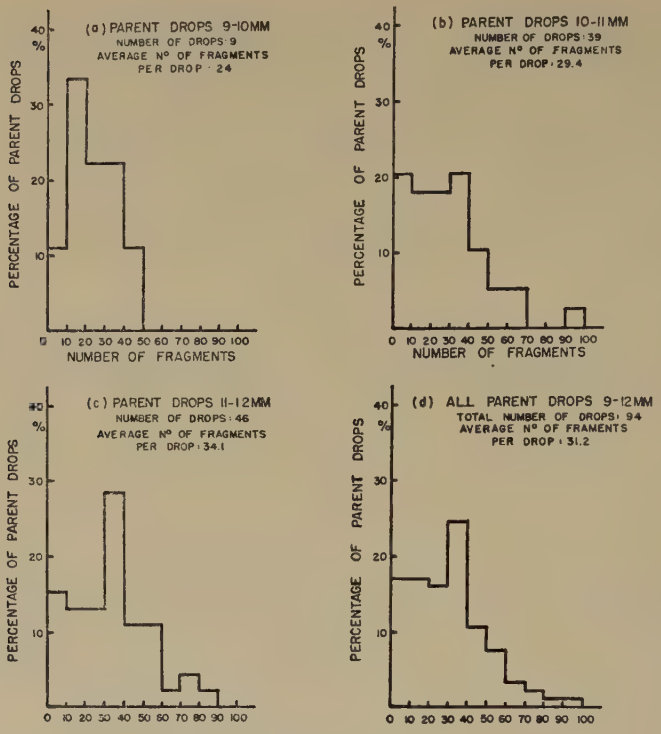
(a) PARENT DROPS 9-10MM
NUMBER OF DROPS: 9
AVERAGE N° OF FRAGMENTS PER DROP: 24
(b) PARENT DROPS 10-11MM
NUMBER OF DROPS: 39
AVERAGE N° OF FRAGMENTS PER DROP: 29.4
(c) PARENT DROPS 11-12MM
NUMBER OF DROPS: 46
AVERAGE N° OF FRAGMENTS PER DROP: 34.1
(d) ALL PARENT DROPS 9-12MM
TOTAL NUMBER OF DROPS: 94
AVERAGE N° OF FRAMENTS PER DROP: 31.2
PERCENTAGE OF PARENT DROPS
NUMBER OF FRAGMENTS

Figure 1. Numbers of fragments.

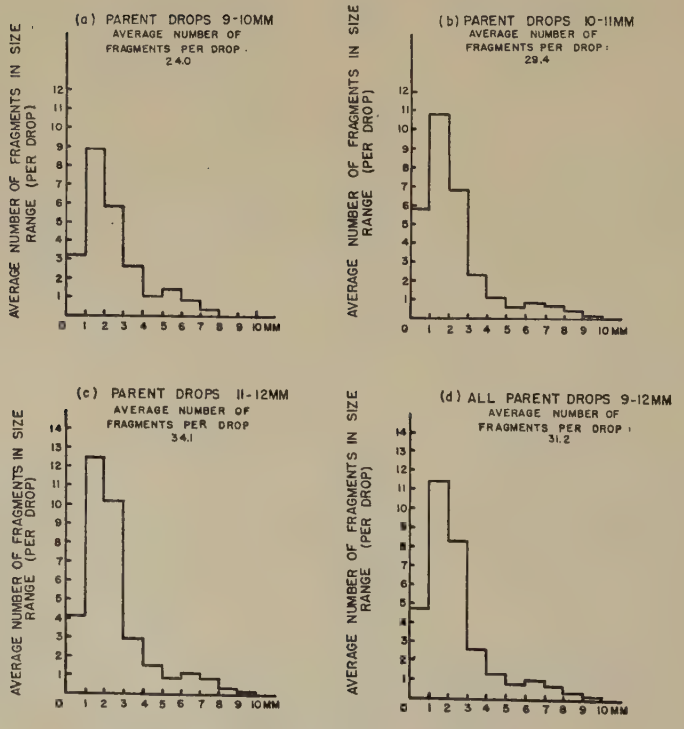
(a) PARENT DROPS 9-10MM
AVERAGE NUMBER OF FRAGMENTS PER DROP: 24.0
(b) PARENT DROPS 10-11MM
AVERAGE NUMBER OF FRAGMENTS PER DROP: 29.4
(c) PARENT DROPS 11-12MM
AVERAGE NUMBER OF FRAGMENTS PER DROP 34.1
(d) ALL PARENT DROPS 9-12MM
AVERAGE NUMBER OF FRAGMENTS PER DROP: 31.2
AVERAGE NUMBER OF FRAGMENTS IN SIZE RANGE (PER DROP)
10MM

Figure 2. Size distribution of fragments.

(d) *Comments*

The critical drop size for break-up, and possibly also the number and size distribution of the fragments, appear to depend greatly on the turbulence regime of the air through which the drop falls. Neither the conditions of this experiment, nor those in previous wind-tunnel experiments, can be expected to reproduce very accurately the conditions in cumulus clouds. In fact it is difficult to see how the latter can be reproduced experimentally even if they are accurately known.

The only immediate possibility of reconciling the results obtained here with those from wind-tunnels appears to lie in a repetition of these free-air experiments, using a tower considerably higher than that of Staff College, Quetta, preferably in an area less subject to earthquakes.

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The authors are indebted to the Commandant of the Command and Staff College, Quetta, for permission to use the College tower for these experiments. We wish to thank also the members of the staff of the Geophysical Observatory, Quetta, and particularly Messrs. Mohd, Shafi Ahmad and A. Haleem, for their ready assistance. This paper is published by permission of UNESCO and of the Director, Pakistan Meteorological Service.

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551.524.32 : 551.524.33 : 517.512.2

Short periods of unseasonal warmth or cold in daily mean maximum temperatures at Bidston

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Summary

Daily means of maximum temperature at Bidston over the 54 years 1900-53 have been examined and in them short periods of unseasonal warmth or cold have been found. Efforts to demonstrate the statistical significance of the anomalies met with difficulties which are outlined, but the results are tabulated and briefly compared with those of previous workers in this field.

1. Introduction

Since Buchan (1867) first propounded his warm and cold periods using observations of daily mean temperatures at five places in Scotland over a period of ten years, considerable attention has been given to the problem of regular and quasi-regular recurrences of weather during the year. Among British workers in this field three have been noteworthy. Brooks (1946) put forward over 20 singularities in our annual march of climate, based on daily pressure values and pressure fields over the British Isles. Lamb (1950), working from 50 years' records of daily weather types affecting the British Isles, was able to propose five climatic seasons and a much greater number of shorter, almost regular, recurrences of characteristic weather. McIntosh

(1953), using 150 years of Edinburgh mean temperatures, and a statistical approach, found that ' the non-seasonal temperature fluctuations are very largely of a random nature but it is highly improbable that they are entirely so.' His results, in the main, did not support Buchan's conclusions regarding Scottish weather, but it should be borne in mind that for some of his results he depended on three-day block means of temperature. This was contrary to one of Buchan's provisos which declared that ' five-day means, and such mean temperatures as have been corrected or manipulated in any way, are unsuitable ' for an investigation into temperature anomalies. Whilst it is not obligatory for a later worker to follow the methods and limitations of a predecessor, it is obvious that in this case failure to do so would reduce the likelihood of finding short-lived weather phases.

The inconclusive nature of our present knowledge of this subject clearly leaves room for further evidence from long-period records. The present paper deals with 54 years' record (1900-53) of daily maximum temperature at the Liverpool Observatory and Tidal Institute, Bidston. The maximum temperature has been preferred for this study to the mean temperature on the following grounds. It is more fundamental, a warm day having a high maximum and a cold day a low maximum for the time of year, whereas mean temperature, though correlating highly with maximum temperature, on many occasions obscures the general weather conditions, as a dull, cloudy day with a small temperature range may easily have almost the same mean temperature as a bright calm day with a high afternoon maximum following a low early morning minimum. Maxima also approximate more closely than means to what the layman understands by warm or cold weather, as he is almost exclusively concerned with daylight hours.

2. Reduction of data

Average maxima for each day of the year (Table 1) were computed and, from these, 31-day means were calculated for 12 evenly-spaced periods, the first centred on 1 Jan., the second on

TABLE 1. Mean daily maximum temperatures for the period 1900-1953

	J	F	M	A	M	J	J	A	S	O	N	D
1	44·4	43·9	45·3	49·7	54·2	62·1	64·9	67·3	63·2	59·0	50·9	46·8
2	45·6	43·9	45·0	49·0	54·8	62·2	65·1	66·0	63·7	58·1	51·3	46·9
3	45·8	44·3	45·7	49·9	55·4	62·2	65·1	65·4	62·4	57·6	51·5	47·3
4	44·4	44·3	45·7	49·7	56·3	63·0	64·1	65·9	63·9	57·9	51·1	46·5
5	44·4	43·6	45·3	49·2	55·8	63·1	65·5	65·4	63·4	58·2	50·1	45·2
6	44·7	44·2	45·4	50·3	56·7	62·9	64·9	66·0	63·2	57·7	50·5	45·8
7	45·1	43·8	46·1	51·2	56·1	62·5	65·0	65·6	63·7	57·4	49·2	45·9
8	44·8	43·6	45·8	50·7	56·6	63·2	65·1	65·1	63·0	57·0	49·1	46·5
9	44·8	44·1	45·2	50·9	56·1	62·7	65·6	65·4	62·4	57·0	49·2	45·9
10	45·1	43·9	45·5	51·0	56·9	62·3	65·5	64·5	62·4	56·7	49·1	45·3
11	45·3	42·6	46·1	52·5	57·4	62·9	66·0	65·2	62·0	56·6	49·6	45·3
12	44·1	42·1	46·4	52·7	57·7	62·6	66·0	65·7	62·1	56·3	49·9	44·9
13	43·7	42·3	46·1	52·5	57·7	62·0	67·7	65·8	61·6	55·4	48·6	45·4
14	44·2	44·1	46·1	52·9	57·4	61·5	67·1	66·4	60·9	55·7	48·2	45·3
15	43·8	44·2	46·8	52·7	56·7	62·2	65·3	65·5	60·0	55·2	47·1	45·5
16	43·7	44·6	47·6	52·3	58·0	63·1	65·4	64·7	60·7	54·7	46·7	45·9
17	44·2	44·6	48·4	51·6	57·9	63·2	66·0	65·2	61·6	55·4	46·8	45·2
18	43·0	44·9	49·0	52·4	58·5	63·4	65·9	65·4	61·4	54·8	46·1	45·1
19	42·8	44·4	47·9	53·0	58·6	63·5	65·1	64·1	60·8	54·7	46·9	44·4
20	43·1	44·0	48·9	53·4	59·4	63·9	66·1	64·3	60·3	54·3	47·7	44·5
21	43·1	43·9	49·4	53·6	60·3	64·1	67·0	64·5	59·2	53·9	47·3	44·3
22	43·2	44·5	48·9	53·2	60·1	63·9	65·3	63·8	59·3	53·0	46·8	44·4
23	43·0	44·5	49·2	53·3	60·6	63·4	65·8	64·2	60·0	52·8	47·2	44·8
24	43·3	44·7	49·0	52·9	61·1	63·0	65·8	64·6	59·6	51·8	47·3	45·0
25	43·3	44·2	48·8	53·4	60·4	62·6	65·6	63·9	59·7	51·3	47·1	45·1
26	43·0	44·3	49·2	53·2	61·6	63·3	65·7	63·9	60·3	50·9	47·0	45·2
27	43·1	44·9	49·2	53·3	61·0	63·7	65·2	64·9	60·3	51·3	47·2	44·8
28	42·7	44·7	49·1	52·1	61·8	64·6	65·6	65·5	60·2	51·4	47·0	44·8
29	42·6	–	49·9	53·2	61·5	64·7	65·2	64·4	59·4	51·6	46·6	44·4
30	43·7	–	48·8	54·2	62·4	64·2	66·1	63·8	58·7	51·7	46·8	44·8
31	43·6	–	49·8	–	63·0	–	66·6	63·3	–	51·7	–	43·7

1 Jan. and so on. These were corrected to 'spot values' for these central days, using the parabolic method due to Brooks and Carruthers (1953), and the spot values were in turn harmonically analysed, giving the results of Table 2 in the form:

$$T_d = A_0 + A_1 \cos\{c\,(d - d_0) - \phi_1\} + A_2 \cos\{2c\,(d - d_0) - \phi_2\} + \ldots . \quad (1)$$

where T_d is the maximum temperature on day d measured from origin d_0 (1 Jan.), A is the amplitude, ϕ is the phase lag and c is a constant converting days to angular measure. This notation has the advantage that ϕ_n/nc measured from 1 Jan. gives the date of maximum.

TABLE 2. RESULTS OF HARMONIC ANALYSIS OF AVERAGE DAILY MAXIMUM TEMPERATURES

	Amplitude (A) (°F)	Phase lag at 1 Jan. (ϕ°)
Mean value (A_0)	54·18	–
First harmonic	11·49	201·0
Second harmonic	0·64	57·8
Third harmonic	0·43	30·8
Fourth harmonic	0·41	39·1
Fifth harmonic	0·30	13·6

TABLE 3. DAILY DEPARTURES OF MEAN MAXIMUM FROM HARMONIC VALUES

	J	F	M	A	M	J	J	A	S	O	N	D
1	0·6	0·4	− 0·1	0·1	− 0·8	1·2	− 0·2	1·1	− 0·2	1·1	− 0·9	0
2	1·8	0·4	− 0·5	− 0·7	− 0·4	1·1	− 0·1	− 0·1	0·4	0·2	− 0·2	0·2
3	2·1	0·7	0·1	0	0	0·9	− 0·2	− 0·7	− 0·7	− 0·1	0·1	0·8
4	0·7	0·7	0	− 0·3	0·7	1·5	− 1·3	− 0·2	0·9	0·4	− 0·1	0·1
5	0·8	− 0·1	− 0·5	− 1·0	0	1·5	0	− 0·6	1·6	1·0	− 0·9	− 1·1
6	1·1	0·5	− 0·5	− 0·1	0·7	1·1	− 0·6	0	0·5	0·7	− 0·3	− 0·4
7	1·5	0	0·1	0·7	− 0·1	0·6	− 0·6	− 0·3	1·2	0·6	− 1·4	− 0·1
8	1·3	− 0·2	− 0·3	0	0·2	1·1	− 0·6	− 0·7	0·6	0·4	− 1·3	0·6
9	1·3	0·2	− 1·0	0	− 0·5	0·4	− 0·1	− 0·4	0·2	0·6	− 1·0	0·1
10	1·7	0	− 0·9	− 0·1	0·1	− 0·1	− 0·3	− 1·2	0·4	0·5	− 0·9	− 0·4
11	1·9	− 1·3	− 0·4	1·2	0·4	0·3	0·1	− 0·5	0·1	0·6	− 0·2	− 0·2
12	0·7	− 1·9	− 0·2	1·3	0·5	− 0·2	0·1	0·1	0·4	0·5	0·2	− 0·5
13	0·3	− 1·8	− 0·6	0·9	0·3	− 0·9	1·7	0·2	0·1	− 0·2	− 0·9	0·1
14	0·8	− 0·2	− 0·8	1·1	− 0·2	− 1·5	1·1	0·9	− 0·4	0·3	− 1·1	0·1
15	0·4	0	− 0·2	0·7	− 0·1	− 1·0	− 0·8	0·1	− 1·2	0	− 2·0	0·4
16	0·4	0·3	0·5	0·1	0·1	− 0·2	− 0·7	− 0·6	− 0·5	− 0·3	− 2·3	0·9
17	0·9	0·2	1·1	− 0·7	− 0·2	− 0·3	− 0·1	0	0·8	0·6	− 2·0	0·3
18	− 0·3	0·5	1·6	− 0·1	0·2	− 0·2	− 0·2	0·3	0·8	0·2	− 2·6	0·3
19	− 0·5	− 0·1	0·3	0·3	0·1	− 0·2	− 1·1	− 0·9	0·4	0·3	− 1·6	− 0·3
20	− 0·2	− 0·6	1·2	0·5	0·7	0	− 0·1	− 0·6	0·1	0·1	− 0·7	− 0·2
21	− 0·2	− 0·7	1·6	0·5	1·4	0·1	0·8	− 0·3	− 0·9	− 0·1	− 0·9	− 0·3
22	− 0·1	− 0·2	0·9	− 0·1	1·0	− 0·2	− 0·9	− 0·9	− 0·6	− 0·8	− 1·2	− 0·1
23	− 0·3	− 0·3	1·1	− 0·2	1·3	− 0·9	− 0·5	− 0·4	0·3	− 0·8	− 0·7	0·4
24	0	− 0·2	0·7	− 0·8	1·6	− 1·4	− 0·5	0·1	0·1	− 1·6	− 0·4	0·7
25	− 0·1	− 0·8	0·4	− 0·5	1·8	− 1·9	− 0·7	− 0·4	0·4	− 1·9	− 0·5	0·8
26	− 0·4	− 0·8	0·6	− 0·8	1·8	− 1·3	− 0·6	− 0·3	1·2	− 2·1	− 0·5	1·0
27	− 0·3	− 0·2	0·5	− 0·9	1·0	− 1·0	− 1·0	0·8	1·4	− 1·5	− 0·1	0·7
28	− 0·7	− 0·5	0·2	− 1·9	1·6	− 0·2	− 0·6	1·5	1·5	− 1·2	− 0·2	0·7
29	− 0·8	–	0·8	− 1·4	1·1	− 0·2	− 1·0	0·6	0·9	− 0·8	− 0·5	0·4
30	0·2	–	− 0·4	− 0·6	1·8	− 0·8	− 0·1	0·1	0·4	− 0·5	− 0·1	0·9
31	0·1	–	0·4	–	2·3	–	0·4	− 0·3	–	− 0·3	–	− 0·2

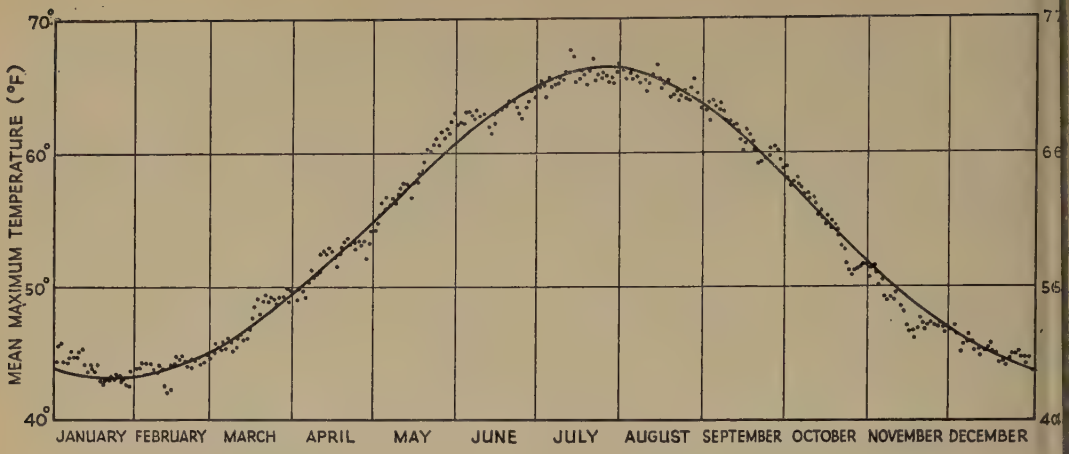

Figure 1. Mean daily maximum temperatures and the corresponding harmonic curve.

Values of daily maxima were then synthesized using the first two harmonic constants only, as, although the amplitudes of the higher harmonics were of the same order of magnitude as the second, the introduction of terms having periods shorter than six months was liable to obscure the anomalies for which the investigation was seeking. Daily departures of the figures of Table 1 from the harmonic values were then extracted (Table 3), and these were found to vary from + 2·3° (mean maximum higher than the harmonic value would indicate) to − 2·6°. The mean departure was, of course, zero, and the standard deviation was 0·81°.

TABLE 4. Warm and cold spells

Date	Mean departure from harmonic values	Remarks
		Warm spells
1-17 Jan.	+ 1·1°	Particularly warm spells on 2 and 3 Jan. (+ 2°) and 6-11 Jan. (+ 1·4°)
16-29 Mar.	+ 0·8°	Particularly warm spell 17-23 Mar. (+ 1·1°)
11-15 Apr.	+ 1·0°	Buchan found cold spell around this date
20 May- 9 June	+ 1·3°	Peak date 31 May (+ 2·3°)
13-14 July	+ 1·4°	Warmest two days of year
27-29 Aug.	+ 1·0°	
4-8 Sept.	+ 1·0°	Peak date 5 Sept. (+ 1·6°)
26 Sept.-1 Oct.	+ 1·1°	Peak date 28 Sept. (+ 1·5°)
5-12 Oct.	+ 0·6°	
24-30 Dec.	+ 0·7°	Peak date 26 Dec. (+ 1·0°)
		Cold spells
11-13 Feb.	− 1·7°	Three coldest days of year
20-26 Feb.	− 0·5°	
9-14 Mar.	− 0·7°	Peak date 9 Mar. (− 1·0°)
24 Apr.-1 May	− 1·0°	Peak dates 28 (− 2·3°) and 29 (− 1·4°) Apr.
13-15 June	− 1·1°	
23-27 June	− 1·3°	Peak date 25 June (− 1·9°)
4-9 July	− 0·6°	Peak date 4 July (− 1·3°)
15-20 July	− 0·5°	Peak date 19 July (− 1·1°)
22-29 July	− 0·7°	
7-11 Aug.	− 0·6°	Peak date 10 Aug. (− 1·2°)
19-23 Aug.	− 0·6°	
22 Oct.- 1 Nov.	− 1·1°	Peak dates 25 (− 1·9°) and 26 (− 2·1°) Oct.
5-10 Nov.	− 1·0°	Peak date 7 Nov. (− 1·4°)
13-23 Nov.	− 1·5°	Peak date 18 Nov. (− 2·6°), largest negative departure of year

3. Discussion

Examination of Table 3 and Fig. 1, in which the mean daily maxima and the harmonic curve of mean maxima are shown graphically, indicates the anomalies of Table 4 to be of interest.

Several difficulties arise when efforts are made to demonstrate the statistical significance of these anomalies.

As the spells have been selected as those dates showing maximum departures from the harmonic values, their significance cannot be shown by any method assuming them to be random departures. In addition the autocorrelation of the figures of Table 3 is + 0·74, and this degree of autocorrelation again renders the normal significance tests inappropriate. This coefficient of autocorrelation shows that the persistence in day-to-day temperature maxima in individual years (computed for two years selected at random as + 0·67 and + 0·81) has been retained in the 54-year means. In view of these considerations and the fact that the temperature series was relatively short for statistical purposes, no satisfactory tests of the significance of the anomalies could be found, but it was felt that the results were worthy of publication.

Brief notes comparing the features of Table 4 with the conclusions of previous workers are given below :

(i) Three of Buchan's cold spells (7-14 Feb., 6-11 Aug. and 6-13 Nov.) are supported, whilst only one, that of 11-14 Apr., is directly contradicted in the Bidston record. No clear evidence relates to the two remaining spells (9-14 May and 29 June-4 July). His warm spell of 12-15 July is directly upheld, whilst the dates of the other warm spells (12-15 Aug. and 3-14 Dec.) show no unseasonal heat or cold.

(ii) Only one of the spells of Table 4 overlaps a change in Lamb's natural seasons, one of the main determinants of which was sudden variation in the frequency distribution of weather types. Many of Lamb's singularities show up very prominently in this Table, in particular his quiet foggy anticyclonic interlude of 15-24 Nov. and his Christmastide thaw and storms of end of year.

(iii) Brooks's singularities are also mainly supported by the Bidston data if the assumption is made that anticyclonic periods are warm, and stormy periods cold, between the middle of March and October, and that the reverse holds during the winter. In general, Brooks's anticyclonic spells are better supported by these data than his stormy periods, probably because anticyclonic spells are more directly allied to temperature extremes than stormy periods, which tend to be very changeable, and variable in temperature.

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Correspondence

551.594.25

The charging of rain by splashing

By J. Alan Chalmers

In a recent paper in this *Journal*, Smith (1955) has suggested that the usual negative potential gradient and positive rain current which are found in quiet rainfall are to be accounted for by the splashing of the rain as it reaches the ground. He supposed the negative charge released to be returned to the earth by point discharge and deduced a relation between potential gradient and rate of rainfall.

There are various observational results which are in disagreement with Smith's suggestion. In the first place, if the observed positive charge on rain is to be ascribed to splashing, then it must be assumed that, when the rain is collected in a well-shielded receiver, as in the measurements of Scrase (1938) and Simpson (1949), the corresponding negative charge is diffusing out of the receiver. Even if this were true in the case of the measurements quoted, it could scarcely have occurred in the measurements of single drop charges by Chalmers and Pasquill (1938) and Hutchinson and Chalmers (1951), where instantaneous charges were recorded, and certainly not in the measurements of Federov (1951) where the measured drop passed right through the apparatus. In all these cases, an excess of positively charged drops was found in quiet rain and this indicates that the drops usually carry a positive charge before they reach the apparatus.

Smith's relation between potential gradient and rate of rainfall is definitely at variance with Simpson's (1949) result, from his Table VI, that there is no such relation. Also Smith's quoted 'usual potential gradient' of -700 v/m differs from Simpson's curve I, Fig. 8, with a most frequent value of -300 to -400 v/m, a value which could not correspond with point discharge at the points in the neighbourhood.

Finally, Smith's theory could not lead to the empirical result of Simpson (1949), that the charge per unit volume on the rain is proportional to the deviation of the potential gradient from the fine-weather value.

In view of these discrepancies, it would seem that Smith's suggestion is inadequate to account for the electrical phenomena associated with quiet rain.

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16 June 1955.

Reply

By L. G. Smith

In general we may assume that some charging is taking place within a cloud from which rain is falling. In the limit this charge will determine the potential gradient at the ground but it is suggested that there is a range of potential gradient in which this effect and the effect of the splashing of drops on the ground are of the same order of magnitude. This is presumably the range which Simpson designates by potential gradients $< |10|$ v/cm. The analysis of the effect of splashing, in the paper under discussion, refers to the particular case of uncharged drops falling from an uncharged cloud, which is not necessarily the most frequent case, and does not attempt to include

y additional effect resulting from charging within the cloud; this is treated separately in the llowing sections of the paper. The complete explanation of observations during quiet rain ust combine these two effects.

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oston 8, Massachusetts, U.S.A.
1 August 1955.

551.551 : 555.558.21

Turbulence generated in standing waves

By R. S. Scorer

In a short contribution to this Journal (**80**, p. 629, 1954, where also symbols are defined) n this topic I used an argument based on perturbation theory which I have since found to be ncorrect. The fallacy arises as follows :

The equation for the horizontal component of motion is

$$(U+u)\frac{\partial u}{\partial x} + w\frac{\partial}{\partial z}(U+u) = -(\theta_0+\theta)\frac{\partial \varpi}{\partial x} \quad . \quad . \quad (1)$$

where u, θ, ϖ are the perturbations of U, θ_0, ϖ_0, and w is the vertical velocity. The perturbation s measured as the difference between the value of a quantity and its value *at the same level* far pstream where the motion is horizontal.

The equation for the slope of the streamlines is

$$\frac{\partial \zeta}{\partial x} = \frac{w}{U+u} \quad . \quad . \quad . \quad . \quad . \quad . \quad . \quad (2)$$

If in this equation and in (1) u and θ are ignored in comparison with U and θ_0, and w is hen eliminated we get

$$U\frac{\partial u}{\partial x} + U\,U'\frac{\partial \zeta}{\partial x} = -\theta_0\frac{\partial \varpi}{\partial x} \quad . \quad . \quad . \quad . \quad . \quad (3)$$

If the motion is periodic then u, ζ, and ϖ are in phase and we can write

$$Uu + U\,U'\,\zeta = -\theta_0\,\varpi \quad . \quad . \quad . \quad . \quad . \quad (4)$$

At an interface between two layers ϖ and ζ are continuous across the interface : Eq. (4) herefore implies that if U' is discontinuous at any level, and θ_0 and U are continuous, then u must be discontinuous. The conclusion is then that waves generate discontinuities of velocity t levels where the velocity gradient in the basic airstream is discontinuous, and this is in con- radiction with the conclusion, Section 6 of my paper in the present issue of this journal (p. 557), hat if the temperature is continuous discontinuities occur at interfaces between two layers only f they already exist in the undisturbed airstream.

When the interface at which there is a discontinuity in U' is displaced to another level in a wave, there is no reason why there should be a discontinuity in velocity gradient at the original evel and so any discontinuity in u' must be equal and opposite to the discontinuity in U'. But in deriving (3) from (1) u' was ignored in comparison with U', the equations being linearized in the normal way. The assumption behind perturbation theory that, provided the perturbation is small enough, terms in the equations involving it to the second degree must be small in com- parison with those involving it to the first degree is not satisfied if a discontinuity is displaced from its original level. As another instance, if there is a discontinuity of density an infinitesimal displacement of it can produce a finite change in the density at a point very close to it; and so (4) cannot be used alone to deduce that discontinuities of u will be generated in waves at discon- tinuities of θ_0, although in fact they are so generated.

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551.515.5

Steering winds for a typhoon moving steadily northward and associated with a well-marked warm sector in the free atmosphere

By H. Arakawa

In the afternoon of 13 Sept. 1954, a mature typhoon ' June ' crossed western Japan showing signs of weakening (Kume 1955). The centre moved steadily northward and *unexpectedly* crossed the east Manchurian coast into eastern Siberia. Typhoons usually turn in a north-easterly direction by the time they cross latitude 35°N. Occasionally, however, continuous movement toward the north are observed. One of the problems in typhoon forecasting is to choose between these alternatives.

The case history of typhoon ' June ' will be used to demonstrate the conditions favourable for northward movement, and also to indicate how the concepts of steering by the upper wind can foreshadow events which are beyond the customary range of forecasts. The data concerning winds aloft at Kimpo Airport, near Seoul, Korea, are analysed and shown in Fig. 1. This time-altitude cross-section shows clearly a fairly strong southerly current up to 20 km, from the early morning of 13 Sept. until 12 hr 15 Sept. (Japanese mean civil time). The upper-air current over Korea are usually from west to east during this season. This strong southerly wind could be considered as an indication of the northward progression of the typhoon ' June.' The upper winds at Kimpo Airport on 16 Sept. show the shifting from south to west. A time-altitude cross-section of the radiosonde data from Kimpo Airport is shown in Fig. 2. The well-marked warm sector corresponds to the strong southerly winds in the high atmosphere. Aerologists have rarely observed pronounced cold fronts such as this one extending up to the 150-mb level.

Figure 1. Time-altitude cross-section of the upper winds over Kimpo Airport (Japanese mean civil time). Arrows fly with the wind (full barb = 10 kt, solid triangle = 50 kt).

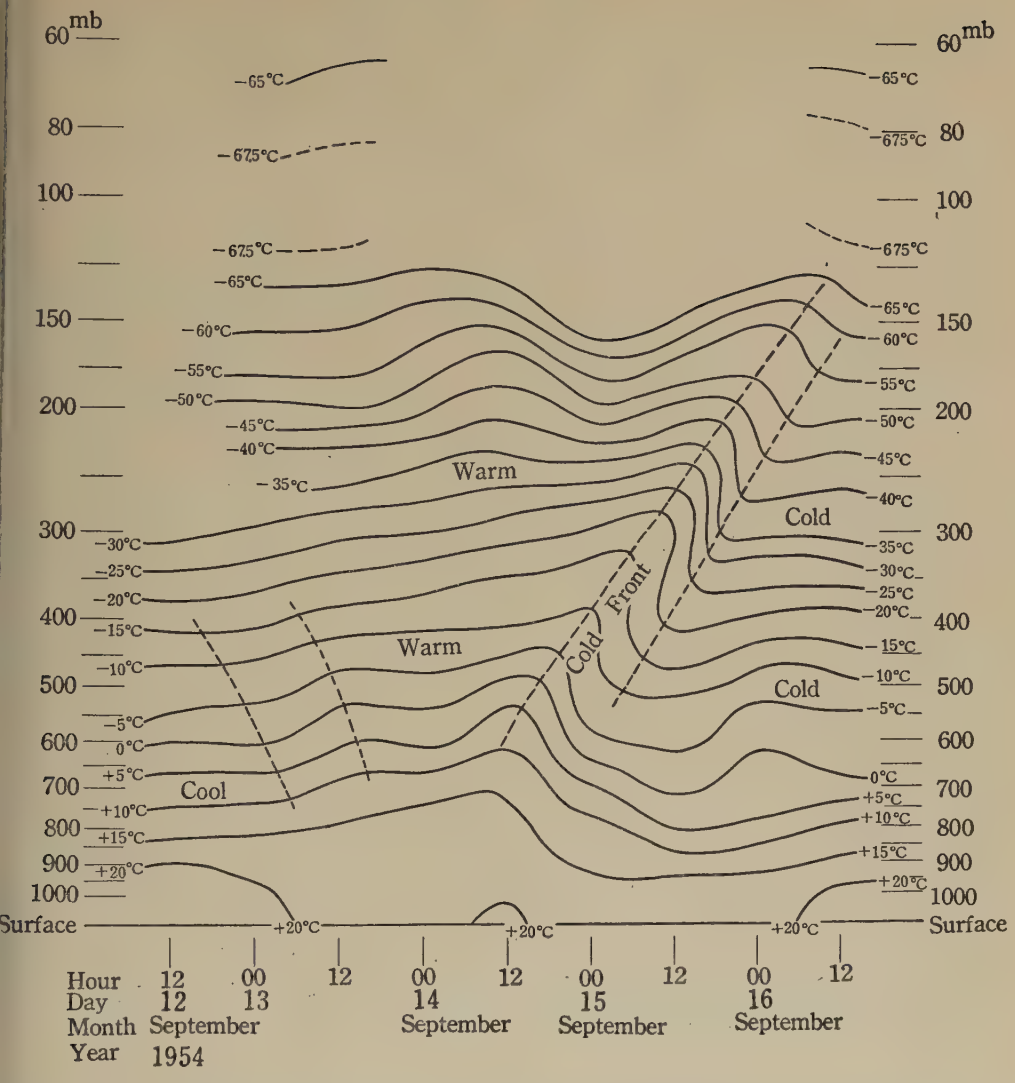

Figure 2. Time-altitude cross-section of the upper air temperatures (Japanese mean civil time). Continuous lines are isotherms.

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6 June 1955.

551.506.3 (4

Constantia Orlebar's weather book, 1786-1808

By Gordon Manley

Early in March 1955 the Society's Honorary Librarian, Dr. G. D. Robinson, made an appea in the form of a letter to *The Times*, for information with regard to the existence of old weath diaries before 1800, and for their loan if possible. Among a number of replies, the ' Weath Book ' kept by Miss Constantia Orlebar of Ecton, Northamptonshire, deserves particular notic The Orlebars of Northamptonshire have held land in that county back to the twelfth centur The Royal Meteorological Society, and the present writer, are most grateful to Captain Richar Orlebar of Hinwick for the generous loan of this weather chronicle, and for some details abo its author.

Constantia Orlebar, born in 1739 at Hinwick House, which is still the home of the hea of the family, was the fifth child of John Orlebar, Clerk of the Privy Council. After her father death in 1765 she moved with her two unmarried sisters to Ecton. They were comfortably c and led an active life, travelling extensively in England and on the Continent, hunting in th winter and frequently visiting London where the family had a town house; incidentally, the correspondence suggests that their dressmakers' bills were quite considerable. Her ' Weath Book ' begins in 1786, by which time she was spending most of the year in or about Northampton shire; and the family history records enough to indicate that in later years she became an invaluab maiden aunt, coming to the rescue when other members of the family were ill, providing a hom for the convalescents, and maintaining a lively interest, by correspondence and otherwise, i affairs great and small.

Like most records of the kind, her notebooks provide interesting testimony to the charact of their author. She clearly resolved to enter, in ink, generally in two lines of her very charmin and elegant script, a description of the weather of each day, in a series of note- or exercise-book bound in contemporary wrapping-paper. Entries are made very systematically with 15 or 1 days to each page, two or three years in each note-book; but the observations for 1800-180 are missing, and in any event were probably much interrupted as she appears to have bee travelling widely about that time. Otherwise, but one day is actually omitted (25 Feb. 1804) and one page, for 17-31 Dec. 1787, is missing.

She had access to a barometer and thermometer, but rarely records their behaviour; an she quite seldom mentions the wind direction, although she has a great deal to say about it characteristics. Some typical entries are :

Jan. 4 1805. Early morning dispersion of fog – and rather more fine but seeming unsettled – with risin wind rough at south east, thawing air, much wasteing snow – hollow wind at night.

Jan. 23 1805. Early morning continuing snow – very cold – rough hollow wind and sleety all day – ver severe weather – night rather turbulent and snowing deep.

Feb. 15 1795. Early Foggy morn – cleared and was fine; thickish sky, yet much sunshine and barometer rising high – Frost at night red clouds at sunset.

Feb. 20 1795. Early Dark and thick clouds – frequent flights of snow – enough to cover the ground – hollow wind, very cold – some snow in the night.

Feb. 24 1795. Early wettish morning – clearer coolish hollow wind – not unpleasant and much less cold still night.

Feb. 27 1786. A misty sprinkling *rain* early – but soon cleared to a spring-like pleasant day – wind rose and was a Blustering cloudy night.

May 27 1793. Early fine but soon arose clouds became very close and must have occasion'd distant Thunder fine setting sun.

June 30 1786. Fine sunshine morn – and not unpleasant clouds.

July 12 1794. Early again very fine and hot – quite oppressing all Day – at night scarcely cooler.

July 24 1786. Fair serene morning – terrific clouds at noon but only a little rain – fine evening but weathe not looking settled.

Aug. 25 1794. Early a hasty shower of rain, soon after the Sun shone hot, but interrupted by clouds an some showers – a fine setting sun.

Sept. 30 1794. Early very hasty rain, changed to be fine – cloudy sunset – wind arose and was a turbulen wet night.

By far the greater part of the record refers to Ecton, a village five miles NE of Northampton (52°15′N, 1°48′W) on rising ground, 300-320 ft above the sea; but for a few weeks each year she was accustomed to pay visits, evidently to friends and relatives, for some days and occasionally longer. The places are named; Hinwick, Northampton, Wellingborough, Haselbech are most often mentioned. Hinwick she visited more often than not in early spring, and again in early autumn; Haselbech about May; London is only mentioned once, for a few days in October 1786. Otherwise her observations refer to the countryside east and north of Northampton within 10-15 mi of her home, and at 200-600 ft.

Her absences from Ecton become less frequent with time, and the last entry in her own hand is for 26 June 1808. From this day onward her sister continues the journal. Constantia's death on 6 July, and her funeral on 11 July, are entered. Then, rather pathetically, her sister resolves to terminate the journal on 31 Dec. 1808 ' on account of the last few days being very dark and my old eyes dim.' One cannot but admire the thoughtfulness and quiet charm of the comments, the steadiness with which they were written on the unlined paper, the evident re-sharpening or perhaps renewal of the quill every two or three weeks, and the maintenance of the observations through the last illness and on the day of the funeral.

Very occasionally an additional comment is made; for example, in June 1804 a note ' the size of bread lessened ' reminds one of the rising prices during the Napoleonic wars. Crops and the flowers receive little mention, although one October day was ' fine for the apple-gathering.' She very generally notes the day near the end of September when fires are first lit, and in the cold September of 1807 ' fire lighted to breakfast ' is entered as early as the 10th. The mail coaches were held up by snow in Feb. 1807. On another snowy day in the village the paths were ' well padded.' Scattered throughout her journal one finds comments on the condition of the roads in winter, an all-important matter in the eighteenth century. Once she makes particular mention of a beautiful light night in January when she was escorted back from a dinner-party. That she was familiar with country weather lore is suggested by her occasional notes on the Saints' days; such comments as ' Fine Barnabas,' ' Bartholomew fine throughout,' occur ; sometimes special note is taken of the weather on the third day of the new moon. ' Moon rather haloish,' ' the sun gilded the yellow leaves,' ' a great rain drop'd the wind,' ' terrific wind, sad Xmas day to my ears,' ' continuing hollow gusts from the settled NE wind,' ' a south wind yet cold,' ' the wind went down with the sun ' make interesting remarks, and from the standpoint of the modern meteorologist the entry for 17 Jan. 1795 is unusually illuminating – ' early snow, the ground cover'd – wind quite subsided – at night kind of rain but very cold – the wind blew by short storms.' Readings of the thermometer are very rarely mentioned, but several appear in the severe January of 1795, the hour not being stated. The lowest was 9° on the 25th (' really pleasant though very cold sitting at church '). Interestingly enough, a few months' observations for the latter part of 1808 on loose paper, in combination with her sister's remarks after Constantia's death, suggest that one of the sisters may have been for a time the keeper of instrumental records; and there is a note of the afternoon reading of 92° on ' Hot Wednesday ' (13 July 1808).

The nature of Constantia Orlebar's entries tells us a good deal; and provides a remarkably interesting sidelight on the make-up of an eighteenth century Englishwoman of considerable social standing, education and accomplishment. She was particularly keenly observant of high-piled cumulus; ' terrific clouds ' often gain mention; and 15 Aug. 1802 is described as ' free of clouds to have apprehensions from.' It becomes very evident that she was particularly sensitive with regard to thunder. Captain Orlebar informs me that this is the more understandable as in a letter written in 1787 she refers to her friend Lady Coventry as ' looking well though her sight much affected through being struck by a flash of lightning on the back of her head as she was sitting in the library last summer.' She notes with evident pleasure how many of the storms avoid Ecton. The village lies with the broad valley of the Nene open to the eastward, along which high-piled cumulus of the hot summer afternoons in the East Midlands is a familiar sight to-day. Because of the obvious care with which she notes ' distant thunder heard,' I have taken out the monthly frequencies below.

She did not relish great heat. On 10 April 1792 ' in a walk we were overpowered by heat ' – this was certainly the first warm day of the year, but at 2 p.m. only gave 64° on a shaded wall in London – and on 28 Aug. 1795, ' autumnal cool evening, very charming to me.' The extracts already given will show her evident interest in the wind, which she disliked when it became too strong and ' tempestous ' (*sic*) or ' turbulent.' Of the latter adjective she is very fond, together with ' terrific,' referring occasionally to wind, more often to large clouds. Expressions such as ' hollow wind at night,' ' russling wind ' (*sic*) and ' whistling hollow wind ' abound and it would be of great interest to know whether the expression ' a hollow wind ' is still used in the

East Midlands and just what it implies. Mr. Charles Linnell, a well-known authority on Bedfordshire and long resident in Pavenham, has no knowledge of it. It might be thought to refer to the noise characteristically associated with the wide chimneys of the past; but what kind of wind makes such a ' hollow, or moaning ' noise ? Does it mean upwards of force 3 ? I have not been able to associate the entries with any particular wind direction, or anything more than the prevalence, or approach, of unsettled weather. It may be noted that Gadbury in his weather journal a century earlier (E. L. Hawke, *Weather*, **4**, 1949) covering the years 1669-89 also uses the term ' great hollow wind.' It seems quite likely to have been a well-known expression in the Midlands that may by now have died out, covering the range force 3-5 or thereabouts. The *New Oxford Dictionary* gives little help; the *English Dialect Dictionary* gives instances of the former use of ' hollow ' in regard to wind in Lincolnshire, Worcestershire and Gloucestershire. I am indebted to Mr. E. L. Hawke for reminding me of Erasmus Darwin's opening line on the signs of rain, ' the hollow winds begin to blow ' (see Inwards's *Weather Lore*, 4th Edition, 1950, p. 227).

No effort is made to reduce the observations, save that after some years she begins to count up the number of days each month on which rain was observed; the totals for the few years that she completed range around 180-190 and stand close to what would be likely to be recorded nowadays under ' days with measurable rain ' in the same district. I have, however, taken out all her observations of snow or sleet (Table 1). Snow was evidently a matter of the utmost interest and Constantia Orlebar was a remarkably keen observer. She was at pains to note any occurrence far into the night and in the early morning, and has many entries such as ' flying snowflakes,' ' a flying sleet ' or ' sleety rains ' which go much further than the majority of the journals of that period. The distribution of her observations by months, and her careful distinction of hail, both suggest that she was an unusually alert recorder of this particular element. They are very well supported by two other contemporary observers nearly equal in keenness, namely Barker of Lyndon (1777-89, 26 mi NNE of Ecton) and Bent of Paternoster Row (1785-1807); while Rooke of Mansfield Woodhouse (1785-1805) who appears to have been consistent, if not quite so alert, gives further corroboration. In as much as she even goes to the trouble to mention, sometimes, a change from rain to sleet in the evening, her total of days probably approaches that which Mr. S. E. Ashmore would regard as ' first-class observation from midnight to midnight ' (*Quart. J. R. Met. Soc.*, **78**, p. 629, 1952). Her average, 26·4 days per annum, is slightly higher than one would nowadays expect from a first-class amateur in the same location and altitude, in excellent agreement with the overall estimates of mean temperature for that period reduced to modern standards (Table 2). For the East Midlands her records provide a useful continuation of those kept by Barker, which end with 1789.

Her specific mention of ' a considerable flight of snow following rain ' on 12 June 1791 appears to be one of the very rare instances of snow falling so far south after the end of May. It receives some corroboration from Bent's journal in London for the same day (' snow said to have fallen on the hills in Sussex '). I have looked up the temperatures and the general meteorological situation for the day in question and they support the likelihood that, with a cold northerly wind and morning temperatures near 40°, snow fell here and there in addition to hail. The lower Lake District hills were white. It is also noteworthy how few occurrences of snowfall were observed in October, and one of them is among the very few ' uncertain ' entries she makes. She has, in addition, much to say of ' rhymy morns ' and ' great fogs,' matters which any East Midland dweller will fully appreciate. Among the pages are a few inserted scraps of paper; on one, she copied out one of the fashionable riddles of the period. Another, dated 1784 and written in a vigorous masculine hand, gives careful instructions how to read the barometer and to use the vernier.

Her sensitivity to thunder, already mentioned, has led me also to take out the monthly and annual averages (Table 3). They correspond fairly with what might by expected to-day, save that June and July may have given more, May and August rather less than we should now observe. There are no present-day stations giving a reasonably long average within twenty miles or so, and the entries are occasionally a little uncertain ; but the overall average of 16·7 days with ' thunder heard ' can be well compared with 13·8 at Cambridge, 19·9 at Belvoir Castle and 22·4 at Bennington (Herts.) given by Marshall (*Quart. J. R. Met. Soc.*, **60**, p. 413, 1934).

As an unusual source of useful meteorological information, and an unusual revelation of a remarkably interesting character, Constantia Orlebar's weather books are indeed welcome. In this age of ' terrific turbulence,' as she might well have called it, far beyond what she could ever have contemplated, twenty-one years of such individually attractive comments remain as an accomplishment which even the hardened physicist can yet esteem.

ECTON AND DISTRICT, SOUTH NORTHAMPTONSHIRE, ABOUT 300 ft

Period 1786-1799, 1802-1808 (21 years).

TABLE 1. AVERAGE NUMBER OF DAYS WITH SNOW OR SLEET OBSERVED TO FALL.

	J	F	M	A	M	J	S	O	N	D	Year
	5·0	5·0	6·1	3·6	0·2	tr.	—	< 0·1	2·0	4·4	26·4
Contemporary records for comparison.											
(a) Bent (Paternoster Row, London) 1786-1807, (b) Hughes (Stroud) 1786-1808, (c) Rooke (Mansfield Woodhouse) 1785-1805. Bent is a ' very good ' record.											
(a)	3·7	3·3	4·7	1·7	0·1	—	—	—	1·2	3·7	18·4
(b)	3·3	2·6	3·0	1·8	0·1	—	—	—	0·8	2·7	14·3
(c)	3·5	3·5	3·7	1·9	0·3	—	—	0·1	0·8	3·4	17·2
Present-day records for comparison.											
(a) Hampstead (450 ft) and (b) Cambridge (41 ft): 1912-1949. (a) is a 'very good' record, (b) 'good standard'											
(a)	5·6	5·3	4·5	2·1	0·2	—	tr.	0·1	1·1	3·4	22·3
(b)	4·1	3·7	3·0	0·9	0·1	—	—	0·1	0·7	1·9	14·5

The probable annual average for the Ecton district by ' very good observation ' over the last 25 years is about 21 days. This is based on 11-14 years at Raunds, Dunstable and Cranfield, extended by comparison with other first-class or very good stations (Cranwell 24·3, Kew 15·6, Oxford 18·7, Mansfield 22·2, Wakefield 27·3, period 1931-54).

TABLE 2. MEAN TEMPERATURE.

November-April for 1786-99, 1801-08	40·4
Present-day, period 1921-1950	41·9

Data from Central England Table, Manley, *Quart. J. R. Met. Soc.*, **79**, p. 257, 1953. These figures lead to an estimated frequency of snowfall 20 to 25 per cent above that of recent years, to which the observations point.

TABLE 3. AVERAGE NUMBER OF DAYS WITH THUNDER HEARD.

	J	F	M	A	M	J	J	A	S	O	N	D	Year
	0·1	0·2	0·4	1·2	2·1	3·9	4·1	2·6	1·1	0·5	0·3	< 0·1	16·7
For comparison (a) Belvoir Castle (35 yr) and (b) Cambridge (40 yr).													
(a)	0·1	0·3	0·7	1·4	3·2	3·9	4·0	3·9	1·3	0·8	0·2	0·1	19·9
(b)	< 0·1	0·1	0·4	0·9	2·2	2·8	2·9	2·7	1·1	0·5	0·1	0·1	13·8

Data from Marshall (*Quart. J. R. Met. Soc.*, **60**, p. 413, 1934).

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6 September 1955.

551.508.94 : 551.594.221

(a) Electrostatic field-changes due to lightning discharges

551.594.221

(b) The development of lightning discharges

By E. T. PIERCE

(Read 18 May, 1955. See *Q.J.*, **81**, pp. 211 and 229)

MR. B. J. MASON : From a study of field-change records Dr. Pierce has been able to deduce something about the charge distribution in a thunderstorm and has obtained some clues on the processes involved in the development of a lightning flash. I feel convinced, however, that further progress in this direction will be rather slow unless we study simultaneously both the electrical and meteorological aspects of the problem. One feels that the distribution of electrical charge should be closely related to the patterns of precipitation and vertical motion, and therefore that it might be very valuable to try and relate the structure and development of the storms as revealed by radar to the electrical characteristics. A study of the field changes suggests that, in the typical South African thunderstorm, the negative charge appears to be distributed throughout a vertical column, whereas in British storms the various charge centres are at much more nearly the same height. We might expect these differences in charge distribution to be reflected in the radar-echo patterns.

The fact that internal flashes and also multiple flashes are more frequent in frontal than in isolated (heat) storms may well be because the former tend to consist of a number of neighbouring centres (cells) of intense precipitation and charge. The average separation of two charge centres involved in successive strokes of a flash to earth, obtained by multiplying the average duration of the *J*-process (·05 sec) by the velocity of the *J*-streamer (3×10^6 cm sec^{-1}), is about 1·5 km; cores of high radar-echo intensity are also separated by distances of this order.

DR. R. H. GOLDE : I should like to congratulate Dr. Pierce on the excellent work presented in these two papers, and I hope that, at some future date, he may be able to supplement his field measurements with at least a few rotating-camera photographs which would help in settling the present uncertainty as to the velocity of the leader stroke. In particular, no evidence seems available for his tacit assumption that this velocity is uniform during the early and later stages of the leader process. Direct photographic evidence on this point would be preferable to comparisons with the long laboratory spark, the velocity of which varies over a wide range according to the external test circuit and which also shows a far from uniform velocity over the distance travelled.

I wonder whether the author can offer an explanation for the exceptionally high current amplitudes discharged in a few positive lightning strokes as evidenced by magnetic link measurements. At the other end of the frequency distribution the large proportion of lightning strokes discharging weak positive currents may be due to the release of electrostatically bound charges during the return-stroke process.

In a recent paper Dr. D. J. Malan (*Ann. Geophys.*, **10**, p. 271, 1954) offers an explanation of the ' S ' field changes based on the assumption of a vertical column of negative charges. The author's comments on this suggestion would be interesting, in particular in view of the predominantly horizontal arrangement of negative charge centres by which thunderclouds in Britain appear to be characterized.

DR. T. W. WORMELL : Dr. Pierce is to be congratulated on the successful completion of this investigation.

Mr. Mason has commented on the finding that, in the majority of occasions, in English storms the change of field observed during the junction-leader process between successive return strokes of a multiple flash to earth is very small, and on the apparent difference between this result and South African observations. It should perhaps be remembered before one bases too

much on this comparison that, as Dr. Schonland has pointed out in correspondence, the South African observations were made at rather smaller distance from the discharge than were the English ones. One would not expect this to affect the argument unless the simple interpretation usually accepted for the cause of the field change during the junction process is wrong. Dr. Pierce's observations, made at considerable distance from the discharge, show that during the junction leader the total change of electric moment is very small where the electric moment includes the charges in the thundercloud itself and any space charges beneath or above it.

I think it can be claimed that we now have a reasonable working picture of the discharge to earth from the instant when the leader process reaches the ground, to the end of the whole process. Certain features of the leader process itself are however still puzzling. What has to happen within the cloud to initiate the leader stroke and to keep it supplied with current? There is a discrepancy between the duration of stepped leaders travelling from cloud to ground, measured by direct photography, and the duration of the initial slow part of the field change, which is an order of magnitude greater, and which Dr. Pierce interprets as entirely due to a downcoming leader. How confident is he of the correctness of this interpretation?

Dr. F. J. Scrase: Dr. Pierce refers to the pockets of positive charge in the base of a thunder cloud as having a trigger effect. Does he regard the existence of such charges as being necessary for a cloud to reach the discharge stage?

I am interested in this point because in 1938 (*Geophys. Mem. Met. Off.*, **9**, No. 75, p. 16), in attempting to explain the observed similarities and differences in the electrification of shower clouds and thunderstorms, I pointed out that the lower region of positive charge which appeared to be characteristic of the active discharge centre of a thunderstorm is absent, or less well developed, in a shower cloud in which no discharges occur. I suggested that a thunderstorm is a further development of a shower cloud in which the stronger vertical air currents cause the development, by the breaking-drop process, of the relatively small regions of positive charge in the lower parts of the cloud, and that unless such charges developed the electric field would not become high enough to initiate discharges.

Dr. C. E. R. Bruce (*communicated*): I am glad that Dr. Pierce has found his observations to be in good agreement with the theory of the leader which envisages transition to arc plus lateral corona currents (Bruce 1941, 1944). Since attempts are still being made to develop the earlier recombination theory of the leader stroke (Loeb 1954, Srivastva and Khastgir 1955), I should like to emphasize once more that the observed values of the pre-discharge fields, the field changes cause by propagation of the leader stroke and the currents in it, differ by factors of the order of 100 from those to be expected on the earlier recombination theory.

As regards the corona constant, k, I would say that my value, of the order of 4×10^{-17}, besides having a certain amount of support from laboratory investigations, also fits into the synthesis which we made of observed data on predischarge fields, field changes, currents in leader strokes and photographic measurements of the velocity of propagation of the leader stroke. (Bruce and Golde 1941). I would suggest that both the value of this corona constant, and the conditions for the onset of arc-like conditions in the discharge channel, are subjects deserving of further theoretical and experimental investigation. I hope that efforts will also be made to investigate the discrepancy between Dr. Pierce's conclusions as to the velocity of propagation of leader strokes and those obtained directly from the photographic investigations of Schonland and McEachron.

References

Bruce, C. E. R.	1941	*Nature*, **147**, p. 805.
	1944	*Proc. Roy. Soc. A.*, **183**, p. 228.
Bruce, C. E. R. and Golde, R. H.	1941	*J. Inst. Elect. Engrs.*, **88**, (II), p. 487.
Loeb, L. B.	1954	*Modern Physics for Engineers*, New York (McGraw-Hill).
Srivastva, C. M. and Khastgir, D. R.	1955	*J. Sci. Ind. Res.*, **14**B, p. 34.

Dr. E. T. Pierce (*in reply*): I entirely agree with Mr. Mason as to the desirability of simultaneous electrical and meteorological studies of thunderstorms. Indeed, for the past few years at Cambridge, we have attempted to synchronize radar and electrical observations, but it has proved difficult, particularly for the rapidly varying conditions characteristic of close storms, to relate the radar structure and the electrical parameters as measured at a single ground station.

I welcome the implication by Mr. Mason that the term ' heat ' storm should be replaced by ' isolated ' storm. ' Heat ' and ' frontal ' were used in my papers purely to facilitate comparison with previous researches. Mr. Mason's suggestion concerning the relative frequency of occurrence of ground, cloud, and multiple flashes, in frontal and isolated storms, seems eminently plausible; as a corollary, it may be pointed out, that, other things being equal, the proportions of cloud and ground flashes should depend primarily upon the height above the earth at which the charge centres originally develop.

Dr. Golde, Dr. Wormell and Dr. Bruce are all concerned with the figure obtained for the velocity of the leader; this figure mainly determines the value derived for k the corona constant. The only ground for the vital assumption that the L field-change is entirely due to a downward leader, is the relatively flimsy evidence of the superimposed radiation pulses (see p. 230 of this volume). Simultaneous photographic and electrical observations, the latter with equipment capable of responding to very low frequencies, are obviously required. In one respect Dr. Golde's remarks need a slight correction; in the corona analysis of the leader field-change, I have used an average value for the leader velocity but I have not assumed that this velocity is uniform; indeed, I have indicated that the velocity depends upon the general field at the leader tip, and will therefore almost certainly diminish as the leader advances.

Regarding Dr. Golde's query about strokes carrying positive electricity to earth, the total resistance of the leader channel when it reaches the ground presumably depends upon its length; the maximum current passing might thus be expected to be very different according to whether the positive charge tapped is at the base or top of the cloud. It would be interesting to know if the charges flowing to earth, for the two types of current, are significantly different.

I am glad Dr. Golde has mentioned the very interesting paper by Malan; this, together with an equally important contribution from Reynolds (*J. Met.*, **12**, p. 1, 1955), became available after the submission for publication of the present papers; it is gratifying to note that many of the experimental results from regions as diverse as England, New Mexico, and S. Africa, are in good agreement. As Dr. Wormell has pointed out in his remarks, it is perhaps premature to assume that S. African and English thunderstorms present, respectively, a vertical and a horizontal arrangement of charge centres. This is certainly an over-simplification for English storms, and there seems no reason why Malan's explanation of the final S field-change should not apply for certain instances in this country. Indeed, my idea of the S (β) change as being due to a gradual neutralization brought about by an upward positive leader, which does not encounter a concentrated negative charge centre, has considerable affinity with that of Malan.

The comments of Dr. Wormell have already been dealt with except for that dealing with the initiation of the leader. It now appears fairly easy to understand how a leader can advance after attaining a length of say 10 m, by which time it in effect produces its own field; it is still obscure, however, how the leader channel first grows from 0 to 10 m. The answer may come from comprehensive laboratory experiments such as those of Loeb (*Phys. Rev.*, **94**, p. 227, 1954).

In answer to Dr. Scrase, I do not think that pockets of positive charge in the base of a thunder-cloud are essential for the development of discharges within the cloud, although I believe that they are probably necessary for the initiation of flashes to earth. The fact that, in the average thunderstorm, flashes within the cloud precede those to ground, is perhaps more in favour of point-discharge from the earth as being the origin of the positive charge pockets, but does not, of course, preclude the operation of such other mechanisms as the breaking-drop process.

Several of the points raised by Dr. Bruce have already been considered. To deal with his other remarks, although I am in agreement with his general concept of arc conditions within the leader, I do not think my observations offer any support to the idea of a glow-arc transition being responsible for the step phenomenon. Furthermore, Professor Loeb in a private communication and also in his most recent publication (loc. cit.), seems to favour the concept that the leader streamer carries a high current and consists of a central core surrounded by a region of less dense space charge; I am accordingly surprised that he is quoted by Dr. Bruce as being opposed to these ideas.

551.501.81 : 551.594.221

Radar echoes from lightning

By R. F. JONES

(Read 18 May 1955. See *Q.J.*, **80**, p. 579)

Mr. B. J. MASON : I think one should be a little cautious in inferring, on the available evidence, that lightning echoes tend to occur outside the regions of strong precipitation echo rather than inside them. This may well be the case on some occasions (e.g. when quasi-horizontal flashes pass between oppositely charged regions inside the cloud). On other occasions, however, might not a transient lightning echo occurring inside the zone of heavy precipitation be masked by the strong precipitation echo ?

Dr. T. W. WORMELL : I am puzzled when reading accounts of radar observations of lightning by the location of the echoing source and by the duration of the echo. Do we really understand in detail what it is we are observing ?

A radar appears, surprisingly, to ' see ' a lightning discharge readily only when it occurs at great heights and outside, although preferably near the edge, of regions of intense precipitation. This appears to be true although the precipitation echo is not saturating the receiver.

Are these fair comments and, if so, has Mr. Jones anything to say concerning them ?

Prof. P. A. SHEPPARD : Is there a consensus of opinion that a cumulonimbus only becomes a thunderstorm *after* the cloud has become sufficiently heavily glaciated and advanced in the life-cycle to display the characteristic anvil of ice ? Mr. Jones's thesis appears to demand this feature. It has seemed to me that the process of charge separation should be well under way when the ice particles (graupel or such like) are still relatively rare compared with the water droplets.

Dr. E. T. PIERCE : Mr. Jones has referred to coincidence of the echoes with the atmospherics received on VHF equipment. Since atmospherics, due to lightning flashes, tend to be less discrete and more prolonged the higher the frequency, I should like to know whether the disturbances on the VHF receiver can be readily identified above the background noise, and if any estimate can be given of their duration.

Mr. R. F. JONES (*in reply*) : It may well be, as Mr. Mason suggests, that a lightning echo occurring within a zone of heavy precipitation will be difficult to detect. Nevertheless for the storms examined, almost every flash which was audible as an atmospheric on R/T was visible on the radar display. An alternative interpretation of this fact, of course, could be that only long quasi-horizontal flashes gave audible atmospherics on R/T, while the flashes to earth passed unnoticed. Dr. Wormell certainly describes the facts as far as the equipment I used was concerned. The height-range aerial used has a relatively wide beam width ($7\frac{1}{2}°$) in the horizontal. Thus a flash propagated almost horizontally could at the ranges used remain within the radar beam throughout its existence and could give a transient echo spreading on the display tube over several miles in range. A vertically propagated flash on the other hand would give a transient echo at one range only and would be less easily noticed when searching for echoes. Thus it is possible that the almost horizontal flashes are most easily picked up and that these occur predominantly in the anvil parts of the clouds.

In reply to Professor Sheppard I think there is a consensus of opinion that the occurrence of lightning is associated with the stages of the cumulonimbus life-cycle after the time when large numbers of ice crystals may be expected to have formed within the cloud (e.g., by penetration above the $-40°$C isotherm).

In reply to Dr. Pierce, there was no difficulty in identifying the typical lightning crackles on the VHF receiver above the background noise. A complete guess as to their duration would be about $\frac{1}{2}$ sec.

Discussion meeting on ' Wind and waves,' 15 June 1955

A joint discussion meeting with the Challenger Society on ' Wind and waves ' was held in the Society's rooms on 15 June 1955. Details of the papers read and the contributions to the general discussion, including those by Dr. H. U. Roll, present as the Society's guest, and by Professor H. Arakawa, who gave a short account of the typhoon of 26 September 1935, are given below.

551.46 : 532.59

(a) The statistical distribution of the heights of sea waves

By M. S. LONGUET-HIGGINS
National Institute of Oceanography, Wormley, Surrey

(See *J. Mar. Res.*, **11**, p. 245)

551.46 : 532.59

(b) The distribution of wave heights and periods in ocean waves

By C. R. DARLINGTON

(See *Q.J.*, **80**, p. 619)

(Both papers were presented together by Dr. M. S. Longuet-Higgins)

Mr. D. E. Cartwright : As a corollary to Dr. Longuet-Higgins's talk, I wish to give an account of the analysis, and application to sea waves, of a probability distribution derived by S. O. Rice in 1944 for the peak values of random noise. This distribution has the advantage over the Rayleigh distribution that it is valid not only for a narrow wave spectrum but for spectra of any width or shape; the appropriate probability curve for a given wave-record is one of a family of curves whose parameter is determined by the shape of the spectrum. It degenerates to the Rayleigh curve only if the spectrum is narrow, but it can also approach the Gaussian curve, and in general is intermediate in shape.

To use this distribution it is necessary to re-define a ' wave-height ' as the height of a crest above, or the depth of a trough below, the mean level. The usual definition, as the vertical distance between a crest and the trough which precedes it, does not lend itself easily to mathematical treatment and, as yet, the distribution of wave heights so defined has been justified only for a narrow spectrum. As re-defined, a wave-height can of course be negative, if a crest occurs below the mean level, or a trough above, and Rice's distribution allows for this.

The assumption made to justify the application of Rice's distribution is very general, namely that the form of the water surface is the linear resultant of a large number of sinusoidal components of different period in random phase relation. It follows that the ordinates and derivatives of the wave surface are distributed normally, and it is then a simple step to derive the distribution of heights of crests and troughs.

It is clear from the slides shown that the theoretical curves give a very good fit to wave heights recently measured in the R.R.S. *Discovery II* and also to some amplitudes of ship motion. It would probably apply equally well to the distributions of slopes, velocities, and accelerations involved, and I would suggest that this probability distribution, being based on very general assumptions, might fit other fluctuating quantities of meteorological interest. A fuller account will be published in due course.

551.551 : 551.554

Some observations of air flow over the sea

By J. S. HAY

(See Q.J., 81, p. 307)

Dr. H. U. Roll : I should like to make two remarks about the variation of shear-stress coefficient with wind speed.

In a former publication (*Geofis. pura appl.*, **21**, 1952) I put together the different values of the shear-stress coefficient resulting from wind-profile measurements of different authors. When computing those values I considered only the middle part of the velocity profiles in order to avoid errors, which might be caused by the influence of stratification and waves. I have now added also the new results of Mr. Hay (see Fig. 1). Apart from a considerable scatter usually connected with such measurements it can be stated that the new results of Mr. Hay are more or less in agreement with the former values. Furthermore, a remarkable feature can be seen : with increasing wind speed there is always, at first, a decrease of the shear-stress coefficient up to a certain wind speed, which may vary between 3 and 6 m/sec. Above this wind speed an increase of the wind stress factor is generally observed. The results of Mr. Hay cover only the latter section.

To give a proper explanation for the observed decrease of the shear-stress coefficient in the region of low winds seems to me very difficult. Since wind-profile measurements made on different sites and with different equipment give similar results, the decrease should probably be considered a real one (assuming of course that it is possible to interpret wind profiles over

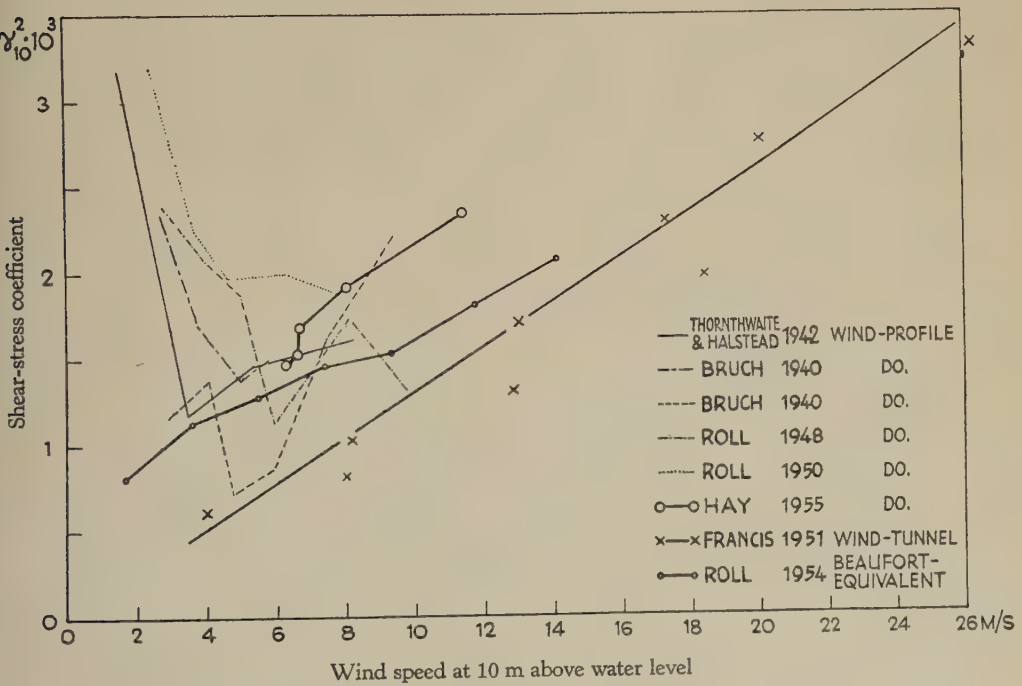


Figure 1. Variation of the shear-stress coefficient with wind speed.

the sea in terms of the theory of the flow in pipes). On the other hand, wind profiles over small ponds at very low winds and with nearly no waves indicate that the wind-stress factor tends to zero as is to be expected. Therefore the observed decrease of the shear-stress coefficient with increasing wind speeds seems to be questionable. More exact and more suitable measurements – especially of the air flow around moving water waves – are necessary to make this point clear. A better interpretation of the observed facts might also be gained by graphing shear-stress coefficient against Reynolds's number, though some difficulty may arise in defining a correct and suitable characteristic length.

More recently I tried to approach the same problem in a different way (*Ann. Met.*, **6**, p. 193, 1953-54). I studied the influence of thermal stratification on the equivalents in knots for Beaufort's wind scale. The result was that equivalent wind-speed values (measured at 19 m a.s.l.) grow if one proceeds from unstable to stable lapse rates. This effect is caused by the influence of thermal stratification on the wind profile. If one considers the Beaufort-number as representative for the wind speed acting directly on the water surface, it is understandable that the corresponding wind speed in 19 m will be higher in stable cases than in unstable ones. By using wind-profile measurements up to 25 m under known thermal stratification I was able to ' calibrate ' the Beaufort-scale in terms of the wind speed at 0·5 m above water level. There seems to exist a linear relationship between Beaufort-number and wind speed close to the water surface. By the aid of this empirical relation and the Beaufort equivalents mentioned above it was possible to obtain corresponding pairs of wind-speed values at the heights of 0·5 m and 19 m a.s.l. under neutral conditions over the open sea. The wind-stress factors resulting from this investigation increase with the wind speed through the whole observed range and run nearly parallel to the wind-tunnel values of Mr. Francis (see Fig. 1).

Mr. J. Darbyshire made a contribution, basing his remarks on his joint paper with M. Darbyshire, ' Determination of wind stress on the surface of Lough Neagh by measurement of tilt,' published in this volume (p. 333).

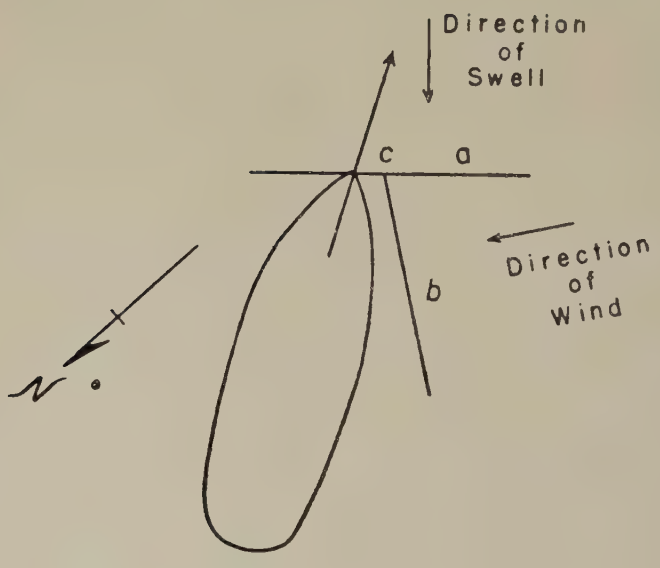
Dr. G. E. R. Deacon : While the agreement between theory and measurement found by Dr. Longuet-Higgins shows that swell, which has a narrow spectrum, behaves as though it had originated at a large number of point disturbances unrelated except in so far as each produces more or less the same range of wavelengths, Mr. Cartwright's success with wider spectra suggests that the same conclusion is reasonable for the generating area itself. If this is so, we shall know a great deal about sea waves as soon as we can find something about one disturbance. While asking the speaker's opinion I feel I have to say what a fine thing it is that oceanographers are at last discussing their work in terms at least as precise as those used by the meteorologists.

Instr. Lt. T. A. Marshall : Instr. Cdr. Darlington has asked me to express his regret at being unable to attend this meeting. In his absence I should like to say that although Dr. Longuet-Higgins's criticism of the relation derived by Cdr. Darlington between T_m and $T_{\frac{1}{3}}$ is justified, the physical unreality of the expression was appreciated by the author, who comments on this point in his paper, and states that the equation cannot be used for extrapolating beyond the ranges of values from which it is derived.

Mr. J. R. D. Francis : As Dr. Roll has shown, Mr. Hay's values of C are all rather greater than those found by other experimenters. This may be due to reflection of the waves by the large pontoon on which the anemometers were mounted. The standing waves thus caused would have a higher relative windspeed over them compared with the usual progressive waves in the sea, and would presumably cause a larger drag on the air above. It would be interesting to see if all experiments with a large pontoon-mounted mast give C values larger than those with a mast from the sea bed unencumbered by a wave-reflecting body.

Prof. H. Arakawa : A main squadron of the Imperial Japanese Fleet was conducting manoeuvres over the sea to the east of Japan on 26 Sept. 1935. A typhoon centre, whose progressive movement northwards was *very large*, passed over this squadron, and the network of meteorological observations, in the storm area of this typhoon in a mature stage, was fairly dense. Composite charts of this typhoon are provided in Figs. 1 (a) and 1 (b). Hourly positions of the main squadron, the torpedo-boat flotilla, the submarine flotilla, the transport ship *Tsurumi*, and the destroyer *Hanekaze* relative to the storm are shown in Fig. 1 (a), in which wind directions are given by small arrows, and mean (over 20-min periods) wind speeds are given for each 5 m/sec. The streamline patterns over the same area, shown in Fig. 1 (b), exhibit a pronounced similarity to Shaw's representation of the air currents in a cyclone. It suggests the formation of fronts in the vortex as it moved into middle latitudes.

(a)



(b)

Figure 2. (a) Photograph and (b) Schematic diagram of the behaviour of the sea in the right rear quadrant (1520 JMT) of the typhoon of 26 September 1935. Based on photographs by Lieut. G. Matsuura, looking to the windward on board the cruiser *Nachi*. The direction of movement of the cruiser was 15 on the 36-point direction scale. a indicates the wave front of swells from the south-east; b, the wave front of wind waves under the influence of wind of speed 27 m sec^{-1} and direction 21; c, pyramidal waves. Note the deviation of the wind to the right of the swell.

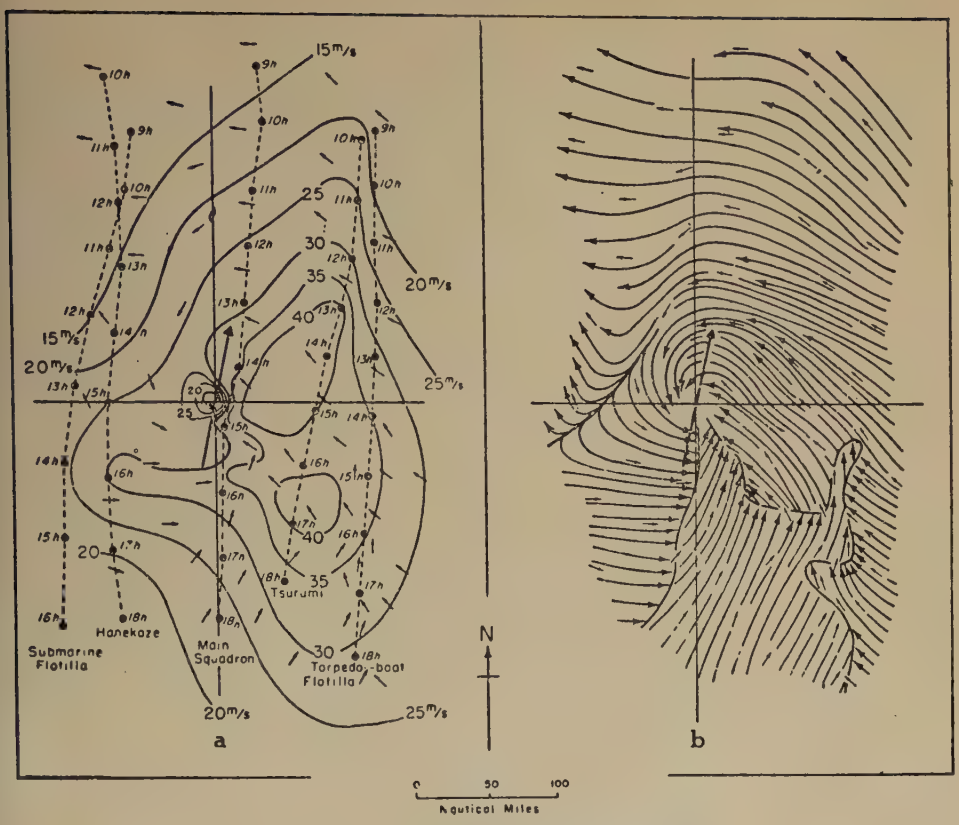


Figure 1. Composite charts of the typhoon, 26 September 1935. Intersection of cross-lines represents the centre of pressure symmetry. The heavy arrow through the centre indicates the direction of movement of the typhoon. (a) Chart of wind speeds and directions in the typhoon. Small arrows represent the wind directions : isovels (lines of constant wind speeds) are drawn for each 5 m sec^{-1}. Dots are positions of naval units at hourly intervals (JMT). (b) Streamline patterns over the same area. (Figs. 1 and 2 are reproduced by courtesy of the Department of Commerce, Weather Bureau, U.S.A., from the *Mon. Weather Rev.*, Vol. **81**, p. 31, 1953.)

The state of the pyramidal sea in the right rear quadrant was photographed by Lieut. G. Matsuura, looking to the windward on board the cruiser *Nachi* (10,000 tons). As illustrated in Fig. 2 (a) and (b) (Plate XI), crossing of two wave fronts, i.e. the front of swell from the south-east, *a*, and the front of wind waves under the influence of wind of speed 27 m/sec from south-south-west, *b*, makes the huge, pyramidal waves, *c*, which are quite dangerous even for battleships.

The wind waves, as related to the movement of the typhoon, during the passage of the storm were also observed. Classical theory on surface waves indicates that wavelength L and period T for deep-water waves are interrelated by the formula $L = 1{\cdot}56\,T^2$, where L is measured in metres, and T in seconds. Measured values for the cruiser *Mikuma* (8,500 tons) were $\bar{L} = 180$ m, $\bar{T} = 13{\cdot}0$ sec at 1250 JMT, and $\bar{L} = 200$ m, $\bar{T} = 13{\cdot}3$ sec at 1445 JMT. Comparisons of measured and computed values for this case gave rather unsatisfactory results after the classical theory, while Neumann-Pierson's spectrum theory gives the relation $\bar{L} = \frac{2}{3} \times 1{\cdot}56\,\bar{T}^2$, which could be applied for such extremely stormy conditions with fairly satisfactory accuracy.

(For a more detailed account see H. Arakawa and K. Suda, 'Analysis of winds, wind waves, and swell over the sea to the east of Japan during the typhoon of 26 September 1935, *Monthly Weather Review*, **81**, p. 31, 1953).

Manchester Meeting, 22-23 July 1955

As reported on page 644 of this issue, two symposia were held at a meeting of the Societ in Manchester on 22 and 23 July 1955. Details of the papers and of the general discussio are given below.

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Medium-scale turbulence in the trade winds

By H. CHARNOCK, J. R. D. FRANCIS and P. A. SHEPPARD
Imperial College, London

SUMMARY

Double-theodolite observations of pilot balloons were made in trade-wind air at Anegad (18°45′N, 64°20′W), a small, low-lying island in the British Virgin Islands, in Spring 1953 Ascents were made at 5 to 15 min intervals over working days of 7 to 8 hr duration on 15 day extending over a total period of 27 days. They provided values of the component air velocitie u, v, w, in orthogonal co-ordinate directions x, y, z respectively (x, y axes fixed in horizonta plane, z vertical), over 50 m intervals of height (20-sec intervals of time) up to about 1·5 km.

Fluctuations in the motion at heights up to 1·5 km and of time-scale varying from abou a minute to several days are analysed and discussed. The work is a pioneer study over mos of this spectrum and has the special interest of dealing with a field of flow in which the direction of the mean surface wind and the thermal wind are opposed (there was a maximum in the profil of mean wind speed at a height of a few hundred metres) and in which cumulus convection is almost always present.

Fluctuations from the mean flow were analysed using averaging periods increasing from about 3 hr up to the whole 27-day period. For none of these averaging periods was there equi partition of eddying energy in the three velocity components; $\overline{w'^2}$ was one to two orders o magnitude lower than $\overline{u'^2}$ and $\overline{v'^2}$, the difference being greater the longer the averaging period.

The covariances $\overline{u'v'}$, $\overline{u'w'}$, $\overline{v'w'}$ were also evaluated for motions at various heights and for the various averaging periods. $\overline{u'v'}$ increased with averaging period and changed sign with height: it appeared that horizontal momentum was being transferred to the north by the larger-scale motions, a smaller amount being transferred to the south by smaller-scale motions $\overline{u'w'}$ and $\overline{v'w'}$ were in general less than $\overline{u'v'}$ and generally varied less systematically with averaging period. Values computed from the smaller components of the motion sampled were in fair agreement with shearing stresses computed by the method of geostrophic departure (using profiles of mean wind and the observed pressure field). The direction of the resultant of $\overline{u'w'}$ and $\overline{v'w'}$ agreed surprisingly closely with the direction of the vertical shear vector of the mean wind velocity, the implied coefficient of eddy viscosity for the spectral range sampled being about 10^5 cm^2 sec^{-1} over the whole range of height.

Dr. F. PASQUILL: In many problems of atmospheric turbulence it is necessary to know something about the ratios of u', v' and w' to the mean speed $\bar{u}$. It would be interesting to see any information which Professor Sheppard can give about these quantities. A further point of some importance is the question of the lapse rate of temperature occurring during the wind measurements which have been described. Presumably there was a close approach to the dry adiabatic lapse rate. Perhaps Professor Sheppard would tell us if there are any data on this point.

Mr. R. J. MURGATROYD: I would like to ask how the shearing stresses measured by Prof. Sheppard compare with those obtained by Bunker from the Catalina aircraft of the Woods Hole Oceanographic Institute which was observing simultaneously in the same area?

Dr. Joanne Malkus : I should like to defer a definite answer to Mr. Murgatroyd's question until Bunker's reduction and analysis of the aircraft data are in a more nearly final form, but certain tentative comments can be made now. The aircraft method of shearing stress measurement gives the contribution of a somewhat different portion of the eddy spectrum from that considered by Professor Sheppard. As Bunker's present results stand, only those eddy sizes between 10 m and about 1,500 m were included. Nevertheless, the shearing stresses he obtained are in the same sense and of the same order of magnitude as those of Sheppard, but numerically less by about a factor of two or somewhat more on occasions. It may be possible from his present records for Bunker to include eddy sizes up to several kilometres, and then the two sets of results should be more comparable.

I should like to comment on the rather low height of the trade-wind maximum in Professor Sheppard's mean wind profile. It is interesting that in a section of the Pacific trade studied by Riehl, Yeh, Malkus and La Seur (*Q.J.*, **77**, p. 598) the average wind maximum occurred at 800 to 900 m, rather than the 300 to 400 m reported by Professor Sheppard. The Pacific section was relatively nearer the northerly fringe of the trades and further downstream, and the season was summer. Professor Sheppard has pointed out that the rate of decrease of wind with height above his maximum was greater, by a factor of two or three, than the mean climatological thermal wind for his area, although he did not have available synoptic data concerning the actual horizontal temperature gradient near Anegada. I have the impression, from numerous photographs of cloud slope and other lines of evidence, that in the Caribbean summer the wind maximum is commonly above cloud base. It would be important to learn whether Professor Sheppard's much lower wind maximum is related to a locally enhanced temperature gradient associated with the gradation from island chain to deep water near Anegada, or whether his measurements are rather more characteristic of winter situations. Although his observations were made in March and early April my impression of the prevailing synoptic picture was that it was more ' wintry ' than normal.

Professor P. A. Sheppard (*in reply*) : The statistics Dr. Pasquill is interested in are given in our complete paper which will shortly be submitted for publication and we would ask to be excused the necessarily lengthy description of them here. We did not make any measurements of the variation of temperature with height but those made by Dr. J. S. Malkus and her group from Woods Hole Oceanographic Institution, using a specially equipped Catalina, were generally consistent with the pattern of cumulus convection described.

We are grateful to Dr. Malkus for answering Mr. Murgatroyd's question about the comparison between our results and those of Bunker; we shall look forward to hearing more details of Bunker's work.

Although Dr. Malkus's earlier work gives the average height of the trade wind maximum (North Pacific, summer) as 800 to 900 m, synoptic data from three Pacific islands (Wake, Johnson, Eniwetok) and from the results of the ' Meteor ' expedition give values for the height of the maximum between 240 and 500 m for winds between 3 and 8 m/sec and between 350 and 800 for winds between 9 and 12 m/sec.

The height is presumably related to latitude as well as to the local geostrophic and thermal wind, but in the absence of an adequate theory of even unaccelerated flow in a baroclinic friction layer one cannot answer Dr. Malkus's question in a satisfying way. We have derived a simple extension to the classical (constant K) theory of the Ekman spiral, and Dr. Ellison is to describe a more complicated one (K linear with height) later today. We hope to compare the wind structure we observed with those predicted on the basis of these assumptions at a later date.

551.511 : 551.557 : 551.547

Geostrophic and ageostrophic wind analysis

By F. B. SMITH

(See *Q.J.*, **81**, p. 403)

No contributions to the discussion following the reading of the above paper were subsequently received for publication.

551.556.4 : 551.510.4

Preliminary studies of the distribution of particles at medium range from a ground-level point source

By F. PASQUILL
C.D.E.E., Ministry of Supply, Porton, nr. Salisbury, and Meteorological Office

SUMMARY

As a basis for planning a comprehensive study of the diffusion of airborne material over distances up to 100 mi a small number of exploratory experiments was carried out during February-April 1955, using as a tracer a fluorescent pigment similar to that described by Braham *et al.* in *Trans. Amer. Geophys. Un.*, December 1952. Samples were taken at distances up to 40 mi downwind of a ground-level source and at heights up to 4,000 ft, using an impactor with a receiving surface in the form of a drum, the instrument being carried across the cloud in a Land-Rover and a Hastings aircraft of the Meteorological Office Research Flight based at R.A.E., Farnborough.

The direction and speed of travel of the clouds were on the average closely in accordance with the measured vector mean wind over the first 2,000 or 3,000 ft, individual deviations therefrom being $\pm 5°$ in direction and ± 4 mi hr^{-1} in speed.

The conditions were mainly moderately unstable in the lower layers with overhead inversions at heights ranging from 2,000 to 7,500 ft. Effective extreme cloud widths ranged from about 40° (angle subtended at source) at 5 mi to about 20° at 40 mi. Vertical diffusion in the layer between the ground and the inversion appeared to be an order of magnitude more rapid than in the lower atmosphere in neutral conditions of stability. No decisive evidence was found as regards the effect of an overhead inversion in limiting the extent of vertical diffusion.

The difficulties encountered in obtaining and interpreting the distribution downwind of a source of short duration emphasize that it would be more profitable to use a longer duration source and to carry out the sampling operations away from the leading and trailing edges of the cloud.

Prof. M. J. LIGHTHILL : Dr. Pasquill compares his observations with a formula of Sutton's which is found to underestimate the diffusion. The use of this formula would be appropriate for ordinary turbulent diffusion, but it is likely that a different mechanism of diffusion, described by G. I. Taylor in *Proc. Roy. Soc.* A, **219**, p. 186, 1953 and **223**, p. 446, 1954, is principally responsible for the phenomena which Dr. Pasquill has observed.

Taylor considered the dispersion of foreign matter in turbulent flow through a pipe. He showed that the variation of velocity of a particle was largely attributable to the variation of its position in the pipe, owing to the variation of mean velocity across the pipe. Rapid turbulent diffusion would tend to *limit* rather than increase the dispersion of foreign matter, by causing the average velocity of any one particle while in a given stretch of pipe to be closer to the average for all the particles, if it were not that rapid diffusion also increases the variation of mean velocity in this flow, which has a bigger effect, so that dispersion is greater for rough than for smooth pipes.

Similar considerations apply to atmospheric diffusion (except over rather short distances). It may well be the variations of altitude of a particle, which are principally responsible for its (vector) variation of velocity. This is supported by the observation that the mean velocity of foreign matter is close to the vector mean of the wind velocity as a function of altitude. It would be interesting to see whether Taylor's theory could be extended to this case, perhaps with the assistance of Dr. Ellison's theory of the Ekman spiral.

Note that, in the early stages before each particle has sampled a large range of altitudes, the observed ' diffusion ' would be greater than in the later stages, when Taylor's theory for the pipe indicates that the averaging process comes more fully into operation. Behaviour like this is noticeable in some of Dr. Pasquill's data.

DR. F. PASQUILL (*in reply, communicated*) : The variation of mean wind velocity with height may well have an important bearing on medium-range diffusion. However, the comparison discussed here was actually against data extrapolated from short range measurements in neutral conditions of stability. This was not done with the object of testing a formula, but in order to bring the new data into perspective, and my own view is that the differences indicated are due in the first place to the powerful mixing action of convective motion.

551.554

The Ekman spiral*

By T. H. ELLISON
Clare College, Cambridge

SUMMARY

Previous workers have used an eddy viscosity varying with height linearly or according to a power law, when discussing the variation of the wind between the surface and the height at which it (supposedly) attains the geostrophic value, but they have not used the appropriate boundary conditions to obtain the familiar forms for the velocity profile very near the surface. In conditions of neutral stability this profile is given by

$$\frac{u}{u_*} = \frac{1}{k}\ln(z/z_0) \quad . \quad . \quad . \quad . \quad . \quad (1)$$

where $u_* = (\tau_0/\rho)^{\frac{1}{2}}$ and $k = 0{\cdot}40$. Eq. (1) corresponds to an eddy viscosity, K_M, given by

$$K_M = ku_* z \quad . \quad . \quad . \quad . \quad . \quad (2)$$

If this formula is assumed to hold at all heights, and the geostrophic wind is assumed to be uniform and constant, it can be shown that

$$(u_G - u) - i(v_G - v) = A\,H_0^{(1)}(4ifz/ku_*)^{\frac{1}{2}} \quad . \quad . \quad . \quad (3)$$

where u is the component of the wind in the direction of the surface wind and v that perpendicular to it; the suffix G denotes the geostrophic value and f is the Coriolis parameter.

Since the Hankel function $H_0^{(1)}(i^{\frac{1}{2}}X)$ has a logarithmic form for small values of X, Eq. (3) yields the logarithmic law for small z. A detailed comparison of Eqs. (1) and (3) shows that, given f and k, the whole flow is determined by any two of u_*, z_0 and G (the magnitude of the geostrophic wind). In particular, the height z_h at which the magnitude of the wind attains the geostrophic value and ψ_0 the angle between the surface wind and the isobars can be found. Corresponding values of these quantities are listed in Table 1.

TABLE 1. SOME CORRESPONDING VALUES OF G/z_0, z_h/G, ψ_0 AND $C_D\left(=\frac{2u_*^2}{G^2}\right)$ FOR LATITUDE 55°N

G/z_0 (sec^{-1})	z_h/G (sec)	ψ_0	C_D
$7{\cdot}5 \times 10$	109	10¾°	0·005
$2{\cdot}0 \times 10^3$	80	8°	0·0025
$1{\cdot}8 \times 10^4$	68	6¾°	0·0018
$1{\cdot}5 \times 10^5$	59	6°	0·0014
$3{\cdot}6 \times 10^6$	50	5°	0·0010
$2{\cdot}9 \times 10^7$	45	4¼°	0·0008

When the geostrophic wind is not uniform but varies with height, the flow is considerably modified. If the geostrophic wind varies linearly with height, this theory based on Eq. (2), can readily be extended. The only modification is that (u_G, v_G) are replaced by (U, V) defined by

$$U = u_G + \frac{ku_*}{f}\cdot\frac{dv_G}{dz}, \qquad V = v_G - \frac{ku_*}{f}\cdot\frac{du_G}{dz}.$$

It can be seen that if the thermal wind (i.e., the rate of change of geostrophic wind with height) is in the same direction as the surface wind, its effect is to decrease or even to reverse the cross-isobar flow, while if it is in the opposite direction it increases the cross-isobar flow. This effect has been observed by Sheppard and his collaborators. Numerical estimates of its magnitude suggest that the thermal wind will have a dominating influence in determining the amount of low-level convergence or divergence produced by friction.

* A full account of this work is included in an article by the author on ' Atmospheric turbulence ' in a book entitled *Surveys in mechanics* to be published by the Cambridge University Press.

Dr. F. Pasquill : I am not clear regarding the effect on Dr. Ellison's analysis of one poi concerning the basis of his assumed variation of K with height. The derivation of K from th universal law relating surface shearing stress and wind profile requires that the shearing stre should be constant with height. However, not very much is known about the variation shearing stress in the height range involved, and we are unlikely to be immediately justified assuming constancy above heights of the order of 100 ft. Perhaps this is actually of r consequence in the treatment which Dr. Ellison has given but it would be interesting to have th point clarified.

Prof. P. A. Sheppard : As an exponent of the difference between the operation of intern friction in barotropic and baroclinic flow and of the importance of this difference to an unde standing of the vertical transfer of zonal momentum in the problem of the general circulatic of the atmosphere, I greatly welcome Dr. Ellison's paper. His analytical treatment will provid relations which can be checked against the data where previously only qualitative consideratio could be employed.

I do not think Dr. Ellison needs to stress the limitations of his treatment to the case neutral static stability. The near-surface layer apart, his assumption of a linear increase eddy viscosity with height could effectively comprise the case of vertical instability such fo example as gives rise to active cumulus convection. It would probably be interesting to hav solutions of the equations of motion for other than the linear increase of eddy viscosity K wit height, e.g., K constant over a substantial depth of the troposphere above a kilometre or so. Th might be more representative for climatological purposes.

Dr. T. H. Ellison (*in reply*) : The law of variation of K with height ($K = ku_* z$) w chosen for mathematical convenience. It would correspond to neutral conditions near th ground, and, in absence of a thermal wind, to slightly unstable conditions higher up whe the stress decreases with height. If there were a significant thermal wind in the same directio as the geostrophic wind, there would be a tendency for the stress to increase with height, and s the specified variation of K would correspond to slightly stable conditions above the surface layer

It should be quite easy to obtain solutions of the equations when $K(z)$ has the form suggeste by Professor Sheppard.

We still know very little about the variation of K above the surface layer but it is probabl rather complicated. Thus, the detailed numerical predictions of the theory cannot be expecte to agree with observation; its primary purpose is to show qualitatively the relative importanc of some of the relevant physical processes which must operate in the real atmosphere.

551.556 : 551.46

Wind stress on water : an hypothesis

By W. H. MUNK

(See *Q.J.*, 81, p. 320)

551.554 : 551.465

Wind stress on a water surface

By H. CHARNOCK

National Institute of Oceanography, Wormley, Surrey

SUMMARY

The vertical distribution of horizontal mean wind in the lowest 8 metres over a reservoir (1·6 km × 1 km) has been measured using sensitive anemometers freely exposed from a fixed mast in water 16 m deep, the fetch being more than 1 km. The resulting profiles are closely logarithmic, the small differences being systematic and possibly due to the thermal instability which existed when the measurements were made.

The usual law for wind profiles in neutral stability is

$$\frac{u}{u_*} = \frac{1}{k} \log \frac{z}{z(0)}$$

where u is the wind speed at height z, k is von Kármán's constant, $\log z\,(0)$ the intercept on the $\log z$ axis, and u_* the so-called friction velocity defined by $\tau_0 \equiv \rho u_*^2$, τ_0 being the surface drag and ρ the density of the air.

To characterize the profiles u_*/k, their slope, was plotted in relation to $z\,(0)$, their intercept; this allowed a direct comparison with other profiles, in particular those recently measured in a laboratory channel by Sibul. The agreement was better than expected and indicated that $z\,(0)$ was comparatively independent of fetch and stability but was largely determined by u_*. The relation between u_* and $z\,(0)$ agreed roughly with the simplest non-dimensional relation between them, $gz\,(0)/u_*^2 = \text{constant}$, so that one is led to a generalized wind profile for flow over a water surface

$$\frac{u}{u_*} = \frac{1}{k} \log \frac{gz}{u_*^2} + \text{constant}$$

which specifies the drag, given the wind at one known height. An approximate value of the constant is 12·5.

This expression can be compared with earlier work. The better wind-profile observations show rough agreement; the experimental scatter is necessarily large since a water surface is aerodynamically much smoother than most land surfaces, precision anemometry in difficult circumstances being required to provide sufficiently precise values. Oceanographic measurements of the tilt of water surfaces are in fair agreement at high wind speeds but at low wind speeds the data are conflicting. The early results which imply that the drag-coefficient (u_*^2/u^2) increases with decreasing wind speed in light winds are thought to be in error; some support for this belief comes from recent estimates of drag using a modified ageostrophic technique, which agree roughly among themselves and with the general expression.

Dr. M. S. Longuet-Higgins: At the National Institute of Oceanography we are in the process of making direct measurements of atmospheric pressure close to the sea surface. The apparatus is contained in a freely floating buoy, flat in shape and about 6 ft in diameter. It carries two gyroscopes, to measure the inclination of the surface, an accelerometer with two integrating circuits to measure vertical displacement, and a simplified microbarograph to measure pressure. Although the apparatus will only give the large-scale pressure distribution, we expect it to give valuable information. Several complete records have already been taken, and are now being analysed.

Dr. W. V. R. Malkus: The papers of Munk and Charnock represent to me a happy step to beyond the many purely descriptive studies of waves in recent oceanographic literature. Though these descriptive studies have encouraged new experimental inquiries, they have also restricted the interpretation of waves to a non-mechanistic statistical language. This would be satisfactory if waves were completely random gas molecules, but waves have organization, waves transport momentum and energy, wave fields are basically non-isotropic and, most important, waves are intimately interrelated with the turbulent wind field which creates them. Hence to study a mechanism for driving waves (wind field given, Munk), or to study the changes in the wind field (effect of wave field given by heuristic formulae, Charnock) is indeed a step toward a mechanistic interpretation. One may hope that the next step is a joint theory; wind and waves both free to achieve a jointly stable system. This will of course be a statistical stability, but in most respects far from random.

551.465

Some observations of turbulence near the sea bed in a tidal current

By K. F. BOWDEN

Oceanography Department, University of Liverpool

Summary

Oceanographers are interested in turbulence in the flow of water in the sea or in estuaries because of its influence on the dynamics of water movements and mixing. These processes are more usually studied by their large-scale effects, and detailed investigation of the turbulent fluctuations has been attempted less frequently in oceanography than on the model scale or in the atmosphere. This paper deals with fluctuations in the velocity components. Earlier observations were reviewed by Bowden and Proudman (1949), who also described experiments in the Mersey estuary, using the Doodson current meter. Further observations on similar lines were described by Bowden and Fairbairn (1952a). An attempt to measure the vertical fluctuations as well as the horizontal fluctuations was made by Francis, Stommel, Farmer and Parson (1953) in the Kennebec estuary, Maine.

The observations described in this paper were made using an electromagnetic flow meter, a new type of instrument for this purpose. It was possible to measure the vertical component w, as well as the longitudinal component u, of the fluctuations, and hence the correlation between them. The instrument was designed at the National Institute of Oceanography, and uses alternating current at 50 c/s, passed through a coil, to produce the magnetic field. Two pairs of electrodes are used to measure the two components of flow in the plane perpendicular to the direction of the magnetic field. The measuring unit is only 4 in. in diameter and two such units can be mounted in various positions on a tripod, 6 ft high, which is laid on the sea bed. The potential difference across the electrodes is approximately 50 μv per kt, and is amplified by electronic circuits after induced p.d.'s, other than those due to the flow of water, have been balanced out. The output is recorded photographically.

The observations at sea were made off Red Wharf Bay, Anglesey, in an area from 12 m to 20 m deep, where measurements of tidal currents and bottom friction had been made previously (Bowden and Fairbairn 1952b). The tidal currents at springs reach a velocity of about 2 kt at the surface. The technique employed was to lower the tripod from an anchored boat, a canvas fin keeping it orientated in the direction of the mean current while being lowered. The tripod with the measuring units remained stationary on the bottom while the records were taken and was raised again before the tidal current changed direction.

Each record gave simultaneous traces of two velocity components, either (a) u and w at one level, or (b) u_1 and u_2 at two different levels, or (c) w_1 and w_2 at two different levels. Fluctuations of periods from about a second upwards were apparent in both u and w components, but the longer period fluctuations were more apparent in u. The records were analysed by measuring them at intervals of 1 sec and calculating the correlations by numerical computation, using a total length of either $3\frac{1}{2}$ or $5\frac{1}{4}$ min of record. Some of the results are summarized below :

(a) From a series of records in which the lower unit was at 75 cm above the bottom while the upper unit was in various positions above it, the coefficient of correlation between the u components decreased from 0·77 at 25 cm spacing to 0·37 at 100 cm spacing.

(b) Under the same conditions, the correlation between the w components was 0·54 at 25 cm spacing, 0·33 at 50 cm and below the level of significance at 75 and 100 cm. The root mean square values of the fluctuations in w were about half those in u.

(c) In a series of records at two heights, 75 and 150 cm above the bottom, the coefficient of correlation between u and w varied between $-$ 0·24 and $-$ 0·42 (u being taken as positive in the direction of the mean flow and w positive upwards). The Reynolds stress term $-\rho\overline{uw}$ gave values between 1·4 and 4·1 dyne/cm^2. Comparing these values with the bottom stress as given by $k\rho U_b^2$, where U_b is the mean current at approximately 1 m above the bottom, led to values for the coefficient k between 1·1 and 2·8 $\times$ 10^{-3}, the average, 2 $\times$ 10^{-3}, being of the same order as has been found by other methods.

The results obtained for turbulence in the sea appear to show a number of features in common with those found in the atmosphere or in model experiments on flow in channels. Observations in a tidal current probably make a useful contribution to the general study of turbulence. Compared with the atmosphere, conditions are simpler in that there is a definite upper boundary to the fluid and in there being neutral stability, at least in the cases investigated so far, while the conditions of the experiment are approximately repeated with the tidal period.

I wish to acknowledge the work of Mr. M. J. Tucker, who was mainly responsible for the design of the electromagnetic flow meter, and Mr. L. A. Fairbairn, who has analysed the records.

References

Bowden, K. F. and Proudman, J. 1949 *Proc. Roy. Soc.* A, **199**, p. 311.

Bowden, K. F. and Fairbairn, L. A. 1952a *Phil. Trans.* A, **244**, p. 335.

1952b *Proc. Roy. Soc.* A, **214**, p. 371.

Francis, J. R. D., Stommel, H., Farmer, H. G. and Parson, D., Jr. 1953 Woods Hole Ocean. Inst., Ref. No. 53-22 (unpublished manuscript).

Discussion

Dr. W. Malkus : Professor Bowden's report of these initial measurements in thoroughly mixed tidal flow suggests that comparison with existing measurements in channel flow can soon be made. The laboratory data go to Reynolds number $\simeq 10^6$. Bowden will be able to extend these data to Reynolds numbers $\simeq 10^7$ to 10^8. I wish to suggest several precautions in comparing the controlled laboratory data and data obtained from natural flows. First, to assure that a steady or quasi-steady state is achieved, measurements can be made through a complete tidal cycle. If the mean profiles and turbulent fluctuations are roughly the same on either side of a tidal maximum, this assures a good approximation to the steady state at the maximum. Since the life time of the pertinent elements in the flow (the ' eddies ') is small compared to the rate of change of the mean flow, we can anticipate that the steady state will be achieved during most of the tidal cycle. Secondly, to assure that the change in depth of the bottom around the mean depth and the change in roughness of the bottom around the mean roughness is not too great, measurements can be taken at upstream and cross-stream stations to determine the homogeneity of the flow. Since the spatial extent of the pertinent elements in the flow is no more than several times the mean depth, we can anticipate that a roughness and depth homogeneity of several depth lengths may be adequate. Lastly, to assure that the wave field and any permanent currents do not alter results significantly, measurements under various wave conditions and careful integration to detect any permanent currents might well be made.

Prof. W. H. Munk : Perhaps some of the uncorrelated fluctuations could be associated with long surface waves. For these the u, w correlation would vanish, and the RMS values of u and w at the two elevations should bear a known relation to one another. This might offer a way to check this suggestion.

Dr. M. S. Longuet-Higgins : Would Prof. Bowden say how far the conditions for his measurements (e.g., bottom roughness) were uniform in a horizontal direction, and whether in his opinion the turbulent fluctuations were such as would be obtained under perfectly uniform conditions ?

Prof. K. F. Bowden (*in reply*) : With Dr. Malkus I look forward to being able to compare measurements in tidal flow with those in laboratory channels and agree that the precautions which he suggests are desirable. By the present technique, observations are made within about 2 hr on either side of maximum flood and ebb flow, but the records analysed so far have not been adequate to show whether there is a significant difference in the turbulence on either side of a maximum. Further observations, which have not yet been analysed, should provide enough data to check this point. As Dr. Malkus suggests, one might anticipate that the turbulence, at least near the bottom, may be regarded as in an approximately steady state at most parts of the tidal cycle. Measurements to check the homogeneity of the flow, as he suggests in his second point, and to eliminate effects of waves or permanent currents would be useful in a future programme.

It is possible that some of the uncorrelated fluctuations in u and w are due to long surface waves, as Professor Munk suggests, but this point has not been investigated.

In reply to Dr. Longuet-Higgins, the soundings indicated that the area around each position, within a radius equal to several times the depth, was fairly uniform in depth and roughness (firm sand, with small fragments of shell). The site was chosen, in fact, with the object of measuring turbulence under conditions as uniform as possible in the sea. Until observations have been made at various sites, however, I would hesitate to say how closely the turbulent fluctuations observed were representative of perfectly uniform conditions.

Proceedings at meetings of the Society

' At Home,' 24 May 1955

The Society and Messrs. Muirhead and Co. Ltd. were joint hosts at an informal ' At Home ' held on Empire Day in honour of visiting meteorologists from the Commonwealth.

A talk on ' Meteorology and Agriculture ' was given by Mr. L. P. Smith and the Mufax equipment was demonstrated.

Special General Meeting, 15 June 1955

A Special General Meeting of Fellows was held to elect the Officers and Council for the year 1955-56.

Fellows had been circulated on 12 April with the list of nominations for Officers and Council for the year 1955-56. No other nominations having been since received, the persons so nominated were declared to be elected unopposed in accordance with By-Law 51.

The Officers and Council for 1955-56 were therefore elected as follows :

President

Reginald Cockcroft Sutcliffe, O.B.E., Ph.D.

Vice-Presidents

Charles Kenneth Mackinnon Douglas, O.B.E., B.A.
Professor Gordon Manley, M.A., M.Sc.
Sir Oliver Graham Sutton, C.B.E., D.Sc., F.R.S.
Thomas Wilson Wormell, M.A., Ph.D.

Vice-President for Canada

Donald Patrick McIntyre, Ph.D.

Vice-President for Scotland

Sir Ernest Wedderburn, O.B.E., D.L., M.A., D.Sc., LL.D.

Treasurer

James Strachan Farquharson, M.A., D.Sc.

Secretaries

Howard Latimer Penman, M.Sc., Ph.D., F.Inst.P.
Richard George Veryard, B.Sc.
George David Robinson, Ph.D., F.Inst.P. (Editor).
Arthur George Forsdyke, Ph.D., D.I.C., A.R.C.S. (Librarian).

Councillors

Charles James Boyden, B.A.
George Arthur Bull, B.Sc.
Henry Charnock, M.Sc., D.I.C., A.Inst.P.
James Marston Craddock, M.A.
Thomas Vivian Davies, M.Sc.
Ronald Frith, O.B.E., Ph.D.
Ronald Westlake Gloyne, B.Sc.
Frank Pasquill, D.Sc.
Professor Norman Pye, B.A.
Richard Segar Scorer, M.A., Ph.D.
Charles Edward Wallington, M.Sc.

Ordinary Meeting, 15 June 1955

The following candidates were declared to have been elected that day by the Council :

FELLOWS

DWYER, Leonard Joseph, 71 St. Helens Road, East Hawthorn, Melbourne, Australia.
EVANS, Edward B., A.M.I.C.E., P.O. Box 340, Tanga, Tanganyika Territory.
EVANS, Lt. Cdr. William Mackenzie, B.Sc., R.N., 39 Broom Water, Teddington, Middlesex.
RIGG, John Brodrick, M.A., " Brendon," Wyatts Road, Chorleywood, Herts.

FOREIGN MEMBER

MAUNG, U. Lun, B.Sc., Assistant Director, Burma Meteorological Dept., 619 Merchant Street, Rangoon, Burma.

ASSOCIATE

CLARKE, Reginald Harry, B.A., B.Sc., 123, Mountain View Parade, Rosanna, N.22, Victoria, Australia.

STUDENT MEMBER

FORSHAW, Joseph Brian, 4 Vaughan Road, Birkdale, Southport, Lancashire.

The meeting took the form of a Joint Discussion on ' Wind and waves ' with the Challenger Society. The following papers were read and discussed :

(1) ' The statistical distribution of the heights of sea waves ' by M. S. Longuet-Higgins (*J. Mar. Res.*, **11**, p. 245, 1952).
(2) ' The distribution of wave heights and periods in ocean waves ' by C. R. Darlington (*Q.J.*, **80**, p. 619).

Both the above papers were presented together by Dr. M. S. Longuet-Higgins.

(3) ' Some observations of airflow over the sea ' by J. S. Hay (*Q.J.*, **81**, p. 307).

Contributors to the discussion included Dr. H. U. Roll, Deutscher Wetterdienst Seewetteramt, Hamburg, who was present by invitation as the Society's guest, and Professor H. Arakawa, who gave a short account of the typhoon of 26 September 1935.

Summer Meeting, 6 July 1955

The Summer Meeting of the Society was held on 6 July 1955 at the College of Aeronautics, Cranfield, Beds., by kind permission of the Warden, Mr. T. J. Palmer.

Some 60 Members and guests attended and were given the opportunity of inspecting most of the departments of the College, including a number of most interesting exhibits. At the close of the Meeting a vote of thanks was moved by the Secretary, Mr. R. G. Veryard.

Two-day Summer Meeting, Manchester, 21-23 July 1955

The second two-day Summer Meeting to be arranged by the Society was held at Ashburne Hall, Manchester University, by kind permission of the Warden. The meeting consisted of two scientific symposia.

The first, on ' Some aspects of turbulence in the atmosphere,' was opened by Prof. P. A. Sheppard, who presented some results of the Anegada (W. Indies) expedition in 1953, based on his joint paper with H. Charnock and J. R. D. Francis on ' Medium-scale turbulence in the trade winds.' He was followed by Mr. F. B. Smith on ' Geostrophic and ageostrophic wind analysis ' (*Q.J.*, **81**, p. 403); Dr. F. Pasquill on ' Preliminary studies of the distribution of particles at medium range from a ground-level point source ' and Dr. T. H. Ellison on ' The Ekman spiral.'

The second symposium was on oceanographic topics, to which the following contributed : Prof. K. F. Bowden on ' Measurements of turbulence near the sea bed in a tidal current '; Prof. W. H. Munk on ' Wind stress on water : an hypothesis ' (*Q.J.*, **81**, p. 320), and Mr. H. Charnock on ' Stress of the wind on the sea.'

In addition, a very enjoyable excursion by coach was arranged to Jodrell Bank Experimental Station, by kind invitation of Prof. A. C. B. Lovell. A tour of the station was followed by tea, when the thanks of the Society were expressed by the Editor, Dr. T. W. Wormell.

Council wish to record their sincere gratitude to the Secretary of the Manchester Centre, Dr. Ruth Rogers, who was responsible for the local arrangements.

Report of the Physical Society Conference on the physics of the ionosphere held at the Cavendish Laboratory, Cambridge, September 1954. The Physical Society, 1955. Pp. 406; Figs., Tables. 40s.

This report contains 50 papers in four sections under the headings : The lowest ionosphere; Irregularities and movements in the ionosphere; The ionospheric F2 layer; and The mathematics of wave propagation through the ionosphere. The first paper in each section is a survey with bibliography outlining the present state of knowledge and serves to provide a background to assist the reader to fit the subjects of the subsequent specialized papers into their context and to appreciate more readily their inter-relations. So successfully is the arrangement of papers achieved, that the volume may well be used as a text-book of ionospheric physics; one would wish indeed that it may serve as a model for conference reports, which so often consist of unrelated papers assembled as in an issue of a scientific journal.

It is the second section which is likely to be of most interest to the meteorologist. It begins with an excellent survey describing how ionospheric irregularities and movements may be observed, their nature, and theories proposed to account for their movements; material from the subsequent papers, which raises new issues of importance, is briefly discussed in an addendum. In a paper on the statistics of star scintillation, the author concludes that the irregularities which cause the scintillations are more likely to be situated in the stratosphere and lower ionosphere than, as is generally believed, in the troposphere. The second paper is concerned with the possibility, first suggested by C. T. R. Wilson, that clouds, by discharging upwards, may produce irregularities in the *E* region above them. It is shown that the majority of *E*-region bursts of ionisation, detectable at vertical incidence on frequencies below 36 mc/s, are in fact caused in this manner and that the world-wide cloud system, which maintains the *E* region the world over at a high positive potential with respect to the earth, does exert a control over the number of *E*-region ionisation bursts in any region. These sporadic bursts and their weak residual ionisation are important in that it is shown that they may be capable of supporting communication systems in the lower v.h.f. band over distances up to 2,000 km, irrespective of season and time of day.

The methods devised during recent years for measuring horizontal velocities of peculiarities in the *F2* region are described and the results discussed in a series of papers. These movements of clouds of ionisation may indicate horizontal transport by winds and the observed corrugated structure in the horizontal contours may be evidence of wave-motion in the atmosphere at these great heights. There is a remarkable consistency in the results obtained by the different methods. In high latitudes, drift is mainly eastwards by day and westwards by night at a velocity of the order of 100 m/sec^{-1}, which changes quite suddenly at dawn and sunset during magnetically quiet conditions. There is also a fairly constant component of velocity directed polewards in summer and equatorwards in winter.

The results of a Viking rocket ascent to 219 km during May 1954 are given and discussed. Recordings by an ion mass-spectrograph which operated successfully during this flight apparently indicated a steady decrease in nitrogen content almost to negligible proportions at high levels, but in the light of other observations, e.g., auroral spectra, this result seems very doubtful. A temperature of 900°K at heights between 300 and 400 km was determined by extrapolation.

There is much to interest the meteorologist in this excellent report.

J.P.

Advances in geophysics, **2**. Ed., H. E. Landsberg. New York (Academic Press). 1955. Pp. x, 286; Figs., Tables. $7.50.

The subject matter of the articles in this collection is so varied that the reviewer cannot be expected to be competent to comment on their accuracy; but, since the aim of the book is ' to acquaint specialists with advances in neighbouring sectors,' it is reasonable to judge the book from the impression it makes and the interest it arouses in a worker in a branch of the subject with which it does not deal.

The book opens with an excellent article by J. S. Marshall, W. Hitschfeld and K. L. S. Gunn on 'Advances in radar weather.' Radar has grown to be one of the most powerful and versatile of meteorological instruments. Here the reader can find a very clear and well illustrated account of the results which have been obtained by its use and an introduction to the associated theory. A second article on meteorology is ' Methods of objective weather forecasting ' by I. I. Gringorten. An objective forecast is one made according to a set of rules so that the forecast follows from the initial data without the use of judgment by the forecaster. Such sets of rules are widely used in forecasting such a quantity as minimum temperature, although the forecaster has to judge whether the prevailing conditions are those in which the rules apply. Most of this article is devoted to a rather general discussion of principles and an enumeration of the quantities for which objective forecasts have been made.

The title of the long article by W. J. Pierson, ' Wind-generated gravity waves ' is a little misleading (or, at least, raises hopes which are not fulfilled). The discussion is mainly concerned with the application of random functions to describe the sea-surface. The reviewer would like to see more discussion of the physics of the problem; but cannot blame the author for failing to explain how the wind produces waves, since almost nothing is known about this.

The last two articles relate to the solid part of the earth. ' Geological chronometry by radioactiv methods ' by J. L. Kulp is interesting and scholarly. The important lead and carbon-14 methods are we described with careful accounts of the errors that can arise in their use. In ' Earthquake seismographs an associated instruments,' H. Benioff describes a large number of seismographs, several of which he has develope himself.

The book is well produced and contains few misprints. The reviewer dislikes the system whereb references are listed numerically in the order in which they appear in the text, but that is a small and con troversial point. All meteorologists should find something to interest them in the book, and a few may b interested by all of it.

T.H.E.

Climatic Atlas of the United States. Comp. S. G. Visher. Cambridge, Massachusetts (Harvard University Press), 1954. Distributed in Great Britain by Oxford University Press. Pp. xii, 403; 1,031 Figs 72s.

It is apparent from only a cursory perusal of this atlas that it contains a very great deal of varied and unusual information. There are over 1,000 maps of the United States as a whole, each portraying the dis tribution of a particular feature of the climate. It comes as a surprise that all this has been put togethe in a not unwieldy volume such as can be perused in comfort in an arm-chair.

By far the largest sections of the atlas are those on temperature and precipitation : these claim abou 370 and 420 charts respectively. The remaining charts, rather under 200 in number, portray pressure and winds, sunshine, cloudiness and fog, humidity and evaporation, and miscellaneous features.

With such large numbers of charts all covering the same area the information for the two main elements is presented from many aspects, and it is obviously designed to meet the practical needs of many differen classes of user. There are of course the usual charts of normal and extreme values of the elements; in some cases these are given for the 52 weeks as well as the 12 months of the year, and in addition for various periods related, for example, to the growing of crops. It is impossible in a brief review to describe the charted information adequately but some idea of its scope may be given by the following selection from the more unusual charts.

Date when the probability of freezing throughout the day reaches 10/25/50 per cent (3 charts);

Date when half the normal number of cold degree-days is reached (mid-season date for fuel requirements);

Total precipitation in so wet a year that only one-eighth of the years received more;

Normal dates of first and last snow-covers of various depths.

By contrast with the sections on temperature and rainfall, that devoted to pressure and winds is brief and inadequate. Monthly mean pressure is shown for eight months only, on very small charts, though for two of the months, Jan. and July, it is repeated on charts of somewhat larger scale. There is a curious mixture of units in these charts, inches of mercury being used for the smaller and millimetres for the larger ones. A section on ' Other winds, sea, lake and land breezes, valley winds, squalls, whirlwinds ' is surprisingly trifling in relation to its impressive title, for it consists of only two rather sketchy charts and a diagram.

The charts are nearly all the same size, about 3 × 5 in., and on the same scale, about 1 : 40 million. Each chart of this pattern takes in the whole of the United States. On the same scale the British Isles would barely cover a postage stamp. Obviously therefore the charts can show only the broadest features. They are probably a good enough guide to the climatic advantages or disadvantages of areas comparable in size with the states of the Union, and this, as is indicated in the foreword, is all the atlas sets out to achieve. But in this respect it will be unfavourably compared with the Meteorological Office *Climatological Atlas of the British Isles.*

The American atlas is printed entirely in black and white : the isopleths are bold and the areas between them are clearly distinguished by various types of shading. This clarity, without the use of colours, is however only achieved at the expense of omitting the latitude and longitude grid and any indication of topography; locations can be judged only in relation to the coastlines and state boundaries, and many of the latter are obscured on the maps with complex isopleth patterns.

The atlas is published at a time when the minds of climatologists are much exercised as to the desirable content and format of a climatic atlas of the world as a whole and for the separate continents, and much useful guidance will be extracted from it. Regarded as a pattern, however, this atlas adheres too rigidly to uniformity of chart size and scale. The great diversity as regards the elements charted, desirable perhaps in an atlas for a single compact territory, suggests that it would be all too easy for plans for the preparation of a world atlas to get out of hand. Whether plans for an atlas of such wide scope could reach fruition in the foreseeable future is another question, as it is doubtful whether the necessarily abundant and highly-processed data exist for any other country or territory comparable in size with the United States.

A.G.F.

Climates in miniature. By T. Bedford Franklin. London (Faber), 1955. Pp. 137; 8 Plates, 10 Figs., 11 Tables. 15s.

Dr. Franklin's interest in plants, animals and birds led him, at an early age, to study the climate in which they live and their reactions to it : over the past 30 years he has carried out experiments and made measurements. His instrumental equipment has been simple, consisting mainly of maximum and minimum thermometers and a few home devices such as the cotton cloth used as an evaporimeter by timing its drying. While Dr. Franklin's own observations provide the material for the text of his book he has obviously kept himself informed of the work of others in the same field.

The book is enjoyable to read and is made fascinating by an approach to each part of the subject through the eyes of a naturalist. A hibernating hedgehog leads to an examination of the heat insulating properties of a bed of leaves, an examination made by inserting a thermometer into the wall of the hedgehog's ' house.' Placing cream cheeses a foot underground to ripen starts a study of soil temperature in which conductivity, specific heat, soil type, soil cover, aeration, water-holding capacity and evaporation are all given their due weight. The chapters on soil temperature and mulches are comprehensive and, to a meteorologist, perhaps the most rewarding in the book. Evapo-transpiration, frost hollows, dew, soil moisture, the effects on micro-climate of shelter, light and shade, climate under glass, climate within a crop and finally climate inside a house, all find a place in the book.

The great merit of the book lies in the easy way in which the author guides his readers through complex problems. Usually this is done without loss in accuracy but occasionally Dr. Franklin is led to dogmatize where controversy still prevails and to over-simplify where the matter is abstruse.

This attractive work, which fills a gap in our literature, will be most valuable in introducing the subject to those meteorologists who have not yet made acquaintance with micro-meteorology.

M.T.S.

Theoretische Hydromechanik, Band I. By N. J. Kotschin, I. A. Kibel and N. W. Rose, translated from Russian into German by J. Sauer and K. Krienes. Berlin (Akademie-Verlag), 1954. Pp. xi, 508; 179 Figs., Tables. 48 DM.

This first volume of Theoretical Hydromechanics is likely to be one of the standard text-books in Soviet Universities, and it has therefore been of much interest to the reviewer in view of its possible recommendation to students of hydromechanics in this country. The present volume covers a limited range of existing incompressible-liquid theory and it is best to present here a brief summary of the contents.

Chap. 1 is kinematical, introducing the basic concepts of continuity and vorticity (38 pp.), while the second chapter introduces the equations of motion (35 pp.). Chap. 3 is concerned with hydrostatics, which is now omitted altogether in English text-books and, one feels, could be omitted here without serious loss. Irrotational-liquid theory appears in Chap. 4 (32 pp.), introducing the complex potential for two-dimensional flow, and vortex motion in Chap. 5 (87 pp.) where the classical theorems of Kelvin, Helmholtz and Bjerknes are proved, and this chapter proceeds through rectilinear vortices to the stability of vortex streets. Flow past two-dimensional cylinders and aerofoil theory occupy Chap. 6 (108 pp.) and this includes the Kirchoff theory of discontinuous streaming, discussing in particular, one is gratified to note, Levi-Civita's method for discontinuous streaming past a cylinder with a curved boundary. Chap. 7 deals with three-dimensional irrotational flow (40 pp.) and includes the source and sink distribution method and the well-known ellipsoidal coordinate solutions of the potential problem. The eighth and final chapter deals with wave motion (125 pp.), long waves, surface waves and so on as in our standard works, and ends with a brief account of long-wave theory of an incompressible liquid on the rotating earth, incorporating the Haurwitz solution.

From the above rather long summary of the contents it will be noted that there is little or no difference between this book and the standard treatises which are available here, in fact the English text-books are probably more valuable to the student since they incorporate a considerable number of exercises which frequently expand the material of the chapter. Very little that is new could be incorporated in the first seven chapters, but the opportunity of introducing some new techniques in the chapter on wave motion has not been taken, and one regrets that the whole field of the development of discontinuities (bores and eagres) in shallow-liquid theory is not even mentioned. My main conclusion is that this book is competently written but is not stimulating; in its favour one must mention the frequent occasions in which the analytical solutions of problems are made more alive by numerical presentations of the solutions.

T.V.D.

Heat conduction with engineering, geological and other applications. By L. R. Ingersoll, O. J. Zobel and A. C. Ingersoll. London (Thames and Hudson), 1955. 3rd Edition. Pp. xiii, 325; 53 Figs., Tables. 32s.6d.

In the preface to the original edition of this book (the edition now in review is the third) the authors declare a twofold aim of providing a treatment of heat conduction suitable for the student with limited time and mathematical preparation and of discussing the many practical applications which can be made, especially by the engineer and the geologist.

After an introductory chapter dealing with nomenclature and definitions, and a short chapter in which the Fourier equation is simply derived and briefly discussed, there are three chapters on steady-state and simply-periodic flow of heat. A discussion of Fourier series then leads into increasingly elaborate problems in three separate chapters. Succeeding chapters deal with formation of ice, graphical and approximation methods for treating problems, and measurements of thermal conductivity. Finally, to the previous (McGraw-Hill) edition there are added two new chapters, one on the heat pump and the other on the treatment of analogous problems arising in the movement of water in porous media.

From a general point of view the authors appear substantially to have fulfilled their declared aim. The discussion of specific applications and collections of problems form useful sequels to the various stages in the fundamental treatment. Certain familiar meteorological problems find place in these applications, notably the diurnal and annual temperature waves, while discussions of basic interest to the micrometeorologist occur here and there in the sections dealing with sources and with drying of solids.

The book is nicely produced, with the mathematics well set out and with useful tables of constants and functions collected at the end. Although not written with their specialized interests to the fore, meteorologists will no doubt find in this book much interest in reading and profit in reference.

F.P.

Asia East by South. By J. E. Spencer. London (Chapman & Hall), New York, (John Wiley), 1954. Pp. x, 453; 136 Figs., 5 Tables. 68s.

In this book Mr. Spencer has made a study which, on the whole, is a notable contribution to our knowledge of the cultural geography of South-east Asia. A very wide field of subjects is covered, However, the solitary chapter of interest to readers of this journal deals with the climatological problems of the region in a fashion which can only be described as superficial and inadequate. The reader often gets the impression that the author is unfamiliar with many of the meteorological factors which are responsible for the various tropical climatic regimes which are described. As a result, much of this chapter is diffuse and obscure, words and phrases not in general use find their way into the text while, elsewhere, well-known terms are used with the wrong meaning. Some statements are misleading, if not incorrect, such as the remark ' the most recognizable front shifts south of the Indies during the northern winter,' which, to say the least, fails to take account of any difference between the characteristics of the fronts of temperate and tropical latitudes. Another vague statement follows on the next page where a good many of the Pacific Ocean typhoons, which do not recurve, are alleged to ' sweep somewhat southwards and across Burmo-Malaya.' The 4 maps at the end of this chapter, being without captions and explanations, are meaningless to the reader. In fact this chapter constitutes the chief blemish on the book as a whole, and in the second edition it will need to be re-edited drastically with the help of a meteorologist familiar with the problems of the region.

The section dealing with mineral geography is up-to-date and informative, while the chapter on the geography of health and disease seems an excellent survey for the student. Regional geography is dealt with in an authoritative and masterly fashion, and the author's first-hand experience of the regions he describes has here produced work which should prove of lasting value. For the geographer, in a broad sense, here is a book deserving his close attention and study; albeit the reader wanting an informed synthesis of climatological events in the region must refer elsewhere at this stage.

R.F.M.H.

Wetterkunde-Wetterforschung : Geschichte ihrer Probleme und Erkenntnisse in Dokumenten aus drei Jahrtausenden. By Karl Schneider-Carius. Freiburg-München (Karl Alber), 1955. Pp. xvi, 423; 16 Plates, 6 maps and diagrams. 27.50 DM.

This handsome book, recently added to the publishers' important Sammlung Orbis Academicus series, may be described as an up-to-date Teutonic counterpart of Sir Napier Shaw's first volume in his great *Manual of Meteorology*. The mantle of Hellmann, who died in 1939, appears to have fallen worthily on Prof. Schneider-Carius as Germany's chief historian of weather science. There is much to praise and little to criticize in the work under review, to which the author has brought the thoroughness and unity of purpose characteristic of his race. The five chapters, each extensively sub-divided and annotated, are concerned respectively with (a) the observation of weather in pre-scientific times; (b) the foundations of weather science; (c) the organisation of weather study and the presentation of data; (d) meteorology down to the beginning of the 20th century and early ideas on the dynamics of the air; (e) modern meteorology and the investigation of atmospheric physics. Particularly valuable features of the book are the list of source-references (12 pp.), the bibliography (31 pp.) and the biographical register (16 pp.). There is a short but adequate index.

It is in his greater use of quotation from source that Schneider-Carius's treatment of the material differs principally from Shaw's. Schneider-Carius allows almost every writer known to have materially advanced knowledge or understanding of weather processes during 30 centuries to speak for himself. So far as the reviewer's somewhat rusty acquaintance with the language enables him to judge, the translation into German

of passages chosen from foreign works for inclusion in the book has been done with all competence. One thing stands apart from the otherwise complete ' Germanity,' As a reminder that Latin was formerly the medium of international communication among scientists (but for the spread of Protestantism it might still have been so to-day), Luke Howard's descriptions of the primary cloud forms are given in his original words – e.g., Cumulus : Nubes cumulata, densa, sursum crescens; Cirro-cumulus : Nubeculae densiores, sub-rotundae, et quasi appositae. Incidentally, one of the interesting facts recorded by Schneider-Carius but not by Shaw is that the great French biologist Lamarck (1744-1829) anticipated Howard by a year or so in publishing a system of cloud classification and nomenclature. Lamarck's five categories were : veil clouds, clouds in chequered layers, fleece clouds, piled clouds, clouds in agglomerations or clusters. This promising scheme failed to achieve recognition. Its originator, accused of ' spending much time in fruitless meteorological prediction ' and rudely told by Napoleon Bonaparte at a public gathering that he should stick to natural history, seems to have lost heart and abandoned his pioneer work in the study of weather. Thus while Luke Howard's name is familiar to meteorologists the world over, Lamarck's has fallen into undeserved oblivion outside the ranks of the biological fraternity.

Not all of the excellent photographs which Prof. Schneider-Carius has selected to illustrate his book appear to have much bearing on the adjacent text, and several of them have been reproduced previously in other publications. Plate 9, portraying an example of what look like ' Contessa del Vento ' clouds developed in a local accentuation of turbulence above Alpine peaks, suggests that such formations may have strengthened, if not originated, the popular belief in ' flying saucers.' Misprints and minor errors are rather frequent, especially in the sections relating to the output of modern British meteorologists, but these are trivial flaws in a work which may be commended to all who find interest in the history of weather science and who possess some knowledge of the German tongue.

E.L.H.

Numerical methods. By A. D. Booth. London (Butterworth's Scientific Publications), 1955. Pp. vii, 195; Figs., Tables. 35s.

This book is based on a series of lectures to undergraduate students, and most of the subject matter has already been adequately dealt with in other standard text-books. The author claims in his preface that the main justification for a further book on computation is that as classical methods of hand computation are not necessarily the best methods for automatic digital computers, it is necessary to have a thorough knowledge of the underlying mathematical principles in order to make the most effective use of the new tools. By not emphasizing which methods are best for automatic calculators and by confining himself almost entirely to the mathematical principles of numerical analysis, the author has failed to produce anything novel. Nevertheless, I do not know of any one text-book which contains all the material in this book, and some people may consider it useful as a reference book.

The first five chapters deal with standard finite difference work such as tabulation, difference tables, interpolation, extrapolation, integration and summation of series. The author then deals with ordinary and partial differential equations, simultaneous linear algebraic equations and non-linear algebraic equations.

Chapter 10 is one of the more interesting chapters. It describes how various functions can be approximated to by polynomials, and this device is often very useful when using automatic calculators.

Finally, there is a chapter on Fourier analysis and a chapter on integral equations.

F.H.B.

The elements of probability theory. By H. Cramér. New York (John Wiley & Sons), 1955. Pp. 281; 29 Figs., 6 Tables. 56s.

Professor Cramér's treatise is an extended translation of a Swedish text-book published in 1926, apparently for use in Universities. The work is in three parts, of which the first covers the historical background and the fundamental theorems of mathematical probability; the second part deals with the main probability distributions, including the binomial, the normal, the x^2 and the t distribution. (This section also covers regressions and correlations with two or more variables.) The third part deals with applications : the main practical techniques are described, and there is a chapter on problems of statistical inference. The treatment is clear, detailed and fairly elementary : for proofs of the more difficult theorems the reader is referred to more advanced treatises. The examples are nearly all non-meteorological, and there is no discussion of the problems arising in dealing with serially correlated data.

This is an excellent book from a somewhat academic standpoint, and should provide a university student with valuable background knowledge in a palatable form, but the reader whose main interests are meteorological will probably find that he needs something either more practical or more comprehensive.

J.M.C.

An introduction to stochastic processes. By M. S. Bartlett. Cambridge University Press, 1955. Pp. xiv, 311 15 Figs., 6 Tables. 35s.

Professor Bartlett defines a stochastic process as an actual (e.g. physical) process with some rando element in its structure. Turbulence is the most obvious meteorological example of such a process, thoug there are many synoptic meteorologists, who even in moments other than those of disillusionment, wou regard the definition as applying accurately to meteorology in general.

This book, then, would be expected to contain much of interest to the meteorologist. Its content in fact a good deal wider than might appear from the above definition, embracing as it does a large pa of the theory of probability. Beginning with a short description of the basic elements of the theory, th author proceeds from the problem of the simple random walk through various types of chain process to final two chapters – to a geophysicist the most interesting of the book – in which methods of statistic analysis of stochastic processes and the correlation and harmonic analyses of time series are considered.

The avowed aim of the author is to bring out the methods and applications of the theory which h develops, at the expense of completeness in the theory itself. Possession of this aim has, however, no resulted in his writing a book suitable for the non-specialist, who is likely to find the complicated matho matics heavy going from the outset. This in fact is a book much more suitable for the specialist produce of statistical methods than for the non-specialist user of such methods.

D.H.M.

Exercises in theoretical statistics. By M. G. Kendall. London (Charles Griffin and Co.), 1954. Pp. vi 179. 20s.

The research meteorologist has always been interested in statistics, because his theories must be teste by comparing a set of predicted numerical values with actual values.

Professor M. G. Kendall's excellent set of exercises in theoretical statistics will be of considerable valu to the meteorologist who has already learned some statistics and wishes to test his progress. They are greater value still as a guide to the types of problem which can be handled by statistical theory.

Exercises in distribution theory and sampling form the first two groups of the book; some of the frequenc distributions may appear to be of academic interest only, but my own experience is that the meteorologi can make use of most distributions. The third group of problems deals with statistical relationship, wit many exercises on correlation and regression; the fourth group deals with the important topics of estimatio and inference. The fifth group of exercises is on time series. The correlation in time of meteorologica elements makes it difficult to estimate monthly or annual variability in terms of daily variability; and th search for periodicities in meteorological data is never-ending. The examples on serial correlation an autoregressive series cannot fail to be interesting and stimulating to the meteorologist.

The 400 exercises are well set out in clear type, and nearly 60 pages of the book are given to answer and hints on solutions.

A.F.J.

Obituary Notices

Prof. Dr. E. van EVERDINGEN

Ewoud van Everdingen, Dutch meteorologist of renown, Honorary Member of the Royal Meteorological Society since 1925, died on 17 July 1955 at the age of 82. Van Everdingen was born in 1873 at Delft. Even as a schoolboy he was interested in meteorology. He entered the University of Leiden in 1891, and as a pupil of Kamerlingh Onnes and Lorentz he was soon completely absorbed in practical and theoretical physics. After a few years he became assistant, and then, at the very early age of 27, ' privaat-docent,' which shows that his famous professors had great confidence in his scientific future. His Ph.D. thesis dealt with the Hall effect and the increase of resistance in an electric field.

Van Everdingen officially entered the meteorological field in 1903, when he became divisional director of the Royal Netherlands Meteorological Institute at De Bilt. Only two years later he was promoted director-in-chief. In 1910 he was appointed professor extraordinary of meteorology, climatology and oceanography at the University of Utrecht. Though active in many fields of geophysics, Van Everdingen was in the first place a synoptic meteorologist, always interested in practical forecasting problems. He was actively engaged in forecasting from 1903, a few weeks after his arrival at De Bilt, until 1938, when he retired as director. He was one of those of the old school who had it all at his finger-tips and who could make remarkably good forecasts with the aid of a few reports and his never-failing judgment of the clouds in the sky. Some members of the Society will remember his lecture on Clouds and Weather Forecasting, given in London (1925). His interest in clouds also led him to study meteorological optics. He was not only an excellent observer himself, but he also arranged for a very reliable network of ' halo stations ' in the Netherlands, the observations from which he always carefully studied and described. Textbooks on meteorological optics will contain Van Everdingen's name for years to come. The same interest in clouds brought his attention to the study of cumulo-nimbi and related phenomena (thunderstorm electricity and waterspouts). He developed a theory on the formation of spouts, which for a long time has been the counterpart of Wegener's theory. He can also be considered as one of our very early agricultural meteorologists. His studies on potato-blight are known all over the world.

Van Everdingen's inaugural lecture as professor dealt with ' The third dimension in the Weather.' The Netherlands Meteorological Service was amongst the first to take intensive part in aerological studies (kites, pilot-balloons and the famous aircraft ascents of Soesterberg). After the First World War Van Everdingen played an important role in international meteorology, first as secretary and later as president of the International Meteorological Organization. He presided over the conferences of directors at Copenhagen and Warsaw. His gift for languages was a great help to the Organization in those days when one did not yet work with interpreters and translators. He had a large circle of friends all over the world, and long after his retirement greetings from many international conferences went to him and Mrs. Van Everdingen. Certainly many British meteorologists of the older generation will remember this doyen of meteorology in Holland, who, though living 50 years later than Buys Ballot, was often considered as his direct successor and who also contributed greatly to the development of meteorology in the Netherlands.

W. Bleeker.

Mr. A. J. CONNOR, M.A.

Abraham James Connor, a former President of the Canadian Branch of the Society, died in his 71st year on 17 June 1955, at his home in Toronto.

A. J. Connor was born on 20 October 1884, in County Fermanagh, Northern Ireland. His father, an Anglican clergyman, emigrated to Canada in 1888. Mr. Connor obtained the degree of Master of Arts from the University of Toronto in 1907 and was employed later that year as a meteorologist by the Meteorological Service of Canada.

Originally a forecaster in Toronto, Mr. Connor became interested in climatological work and soon took charge of the Climatological Section of the Canadian Service, a post which he held until his retirement in 1950. During his 42 years in the Service, Mr. Connor published many papers, articles and reports on the climatology of Canada.

Major studies of the temperature and precipitation of Western Canada were undertaken under Mr. Connor's direction and were published from 1915 to 1920 by the Meteorological Service. His articles on the precipitation of Canada, including a study of the ' Droughts in Western Canada,' appeared in the *Canada Year Book* series from 1926 to 1933. During the mid-1930's Mr. Connor wrote the ' Climate of Canada ' for the Köppen-Geiger *Handbook of climatology*; a by-product of this study was the *Climatic maps of North America* folio which he prepared in collaboration with Dr. C. F. Brooks. Another study prepared during this period was the *Climate of Manitoba*, which was published as part of an economic survey report of that province. Mr. Connor's last study of the climate of Canada was published in the *Canada Year Book* in 1949 and is the standard reference on the subject.

Mr. Connor's advice on the meteorology and climate of Northern Canada was always much in demand and the reports of several planning projects include his writings. He was also greatly interested in the weather and climate of British Columbia and in 1949 he supervised the preparation of a booklet, *The frost-free season in British Columbia.*

Along with his climatological work, Mr. Connor maintained an active interest in forecasting and i 1937 became responsible for the Okanagan Valley Frost Warning Service in British Columbia. From th year until his retirement Mr. Connor spent two months each spring as chief forecaster in this special service.

With the formation of the Canadian Branch of the Society in 1940, Mr. Connor became a Fellow an served as President of the Branch in 1948-9. Mr. Connor's vast knowledge was readily drawn upon b all who knew him and his ready wit was enjoyed by all.

ANDREW THOMSON.

MISS C. F. MARSHALL

Intimation has recently reached the Society of the death, in August 1954, of Miss Cordelia France Marshall, a Life Fellow, elected on 15 January 1908. She was the younger daughter of Mr. Stephen A Marshall, J.P., who held Fellowship of the Society more than half a century ago. Mr. Marshall starte observations of rainfall at his home, Skelwith Fold, Ambleside, in 1890, and finding great interest in th notoriously wet climate of the Lake District he gave increasing attention to its study as time went on Eventually he was able to arrange for the establishment of about 25 raingauges in the hills and dales o Cumberland and Westmorland, some of them at remote sites difficult of access. Over a number of year he issued a valuable monthly abstract of his records, copies of which were supplied regularly to this Societ and to the British Rainfall Organization. When her father died, in 1904, Miss Cordelia Marshall fulfille what she knew would have been his wish by maintaining as many of the observations as she could and b sending the returns for publication. Through her persevering efforts several of the gauges were kept in continuous operation for upwards of 35 years after Mr. Marshall's death, until difficulties resulting fro the 1939-45 war intervened. Seldom can two generations of a family have made such important contributio to knowledge of rainfall in any country.

E.L.H.

CAPTAIN T. W. W. PICKARD

We have to record the death, on 18 March 1955, of Captain Thomas William Walter Pickard, a Life Fellow of the Society, elected on 15 November 1911. Captain Pickard had seen service over a long ter of years in the Mercantile Marine as a master mariner. During the latter part of his life he resided a Worthing.

E.L.H.

MR. W. R. PENNELL

Walter Richard Pennell, a Fellow of the Society since 1930, died on 6 July 1955 in his 91st year. He was well known in Lincolnshire as a nurseryman and seedsman. The close bearing of climate and weathe on his work impressed him from an early age, and during the greater part of his life he was an assiduous student of atmospheric effects on soil and growth. Over a long series of years he undertook regular meteoro- logical records, these being maintained unbroken until within a few days of his death. Since 1924 he had been a contributor to the annual volumes of *British Rainfall*. Mr. Pennell was also greatly interested in the weather-lore of the countryside, and by combining a wide knowledge of natural signs with the indications derived from his daily instrumental observations he achieved considerable success as a local forecaster. We are indebted to his son, Mr. W. Pennell, for much of the information given in this notice.

E.L.H.

Corrigendum to October 1954 issue

K. L. S. Gunn and T. W. R. East, p. 525—

In Table 2, to be consistent with $n = 1{\cdot}78$, the value of $|K|^2$ should be 0·208 and not 0·197, and it should be noted that this value of $|K|^2 = 0{\cdot}208$ is for ice of *unit density*. This revision affects the following :

(1) Eq. (17), factor of first term should be increased by 5·6 per cent from 0·00349 to 0·00368.

(2) Table 4. Second row, snow. Each factor multiplying $R^{1{\cdot}6}$ should be increased by 5·6 per cent.

(3) Table 5. Second, third and fourth row, snow. Each factor multiplying $R^{1{\cdot}6}$ should be increased by 5·6 per cent.

The authors are indebted to Mr. B. J. Mason, whose query brought this error to light.

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See supplementary index below for newly-elected Members

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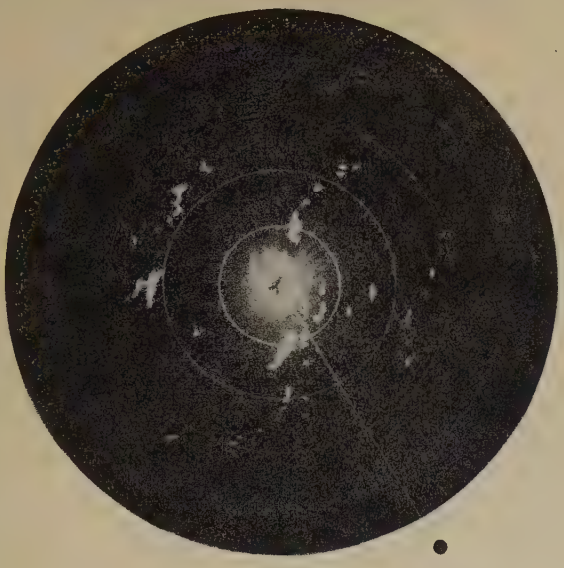

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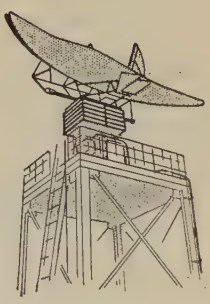

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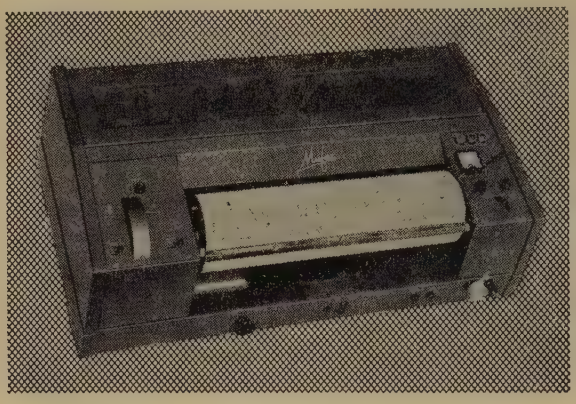

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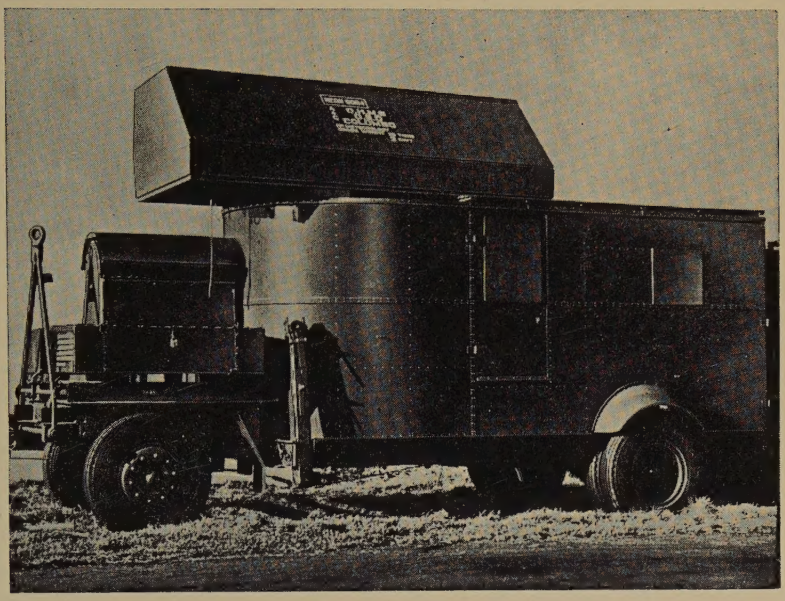

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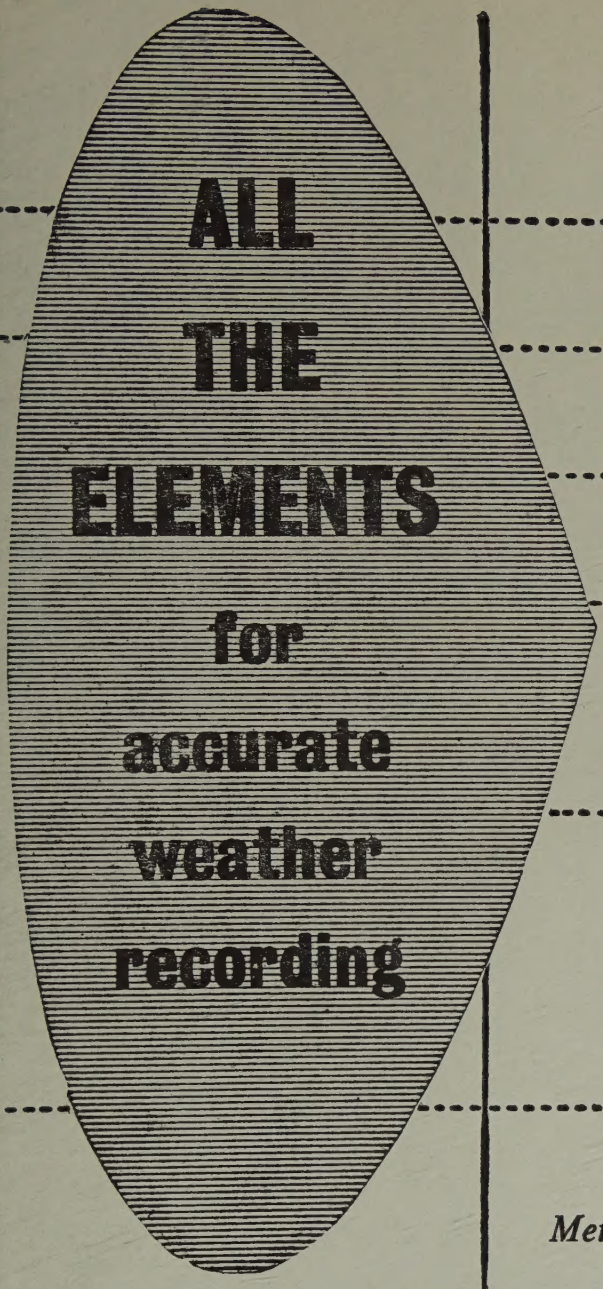

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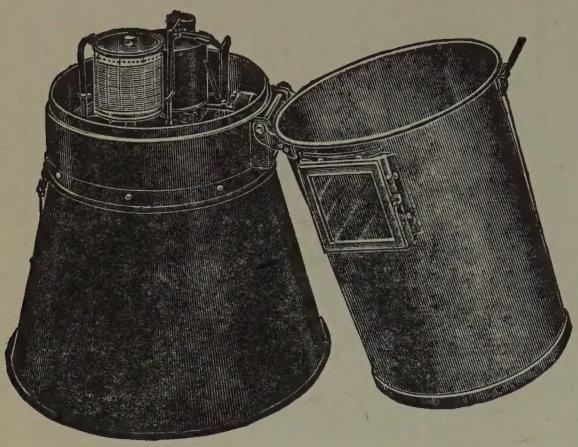

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